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The petrology of the evaporites of the Eskdale no. 2 boring, east Yorkshire.

Part II. *The middle evaporite bed.*

(With Plates IX and X.)

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I. Introduction and acknowledgements.

THE D'Arcy Exploration Company's boring, put down in 1938–39 near Aislaby, Eskdale, passed through three Permian evaporite beds. A petrological account of the rocks of the lower evaporite bed has been given in an earlier communication (Stewart, 1949). The present paper deals with the petrography of the middle evaporite bed. The material examined comprises:

Core samples of known depth collected by I.C.I. while drilling was in progress, and consisting of pieces of core, up to 6 inches in length and 4½ inches in diameter, taken at intervals of 2 feet in the following depth ranges: 3910–22, 3950–66, 3970–4008, 4010–32, 4034–46, 4050–64, 4066–68, 4070–84, 4090–4110, 4118–36, 4138–44, 4146–50, 4176–78 feet below surface.

It will be seen that there is a significant gap from 3922 to 3950 feet, from which no core samples were recovered. The range 3950–4150 feet is fairly well represented, but the lower 46 feet of the bed is represented by only one core sample.

Core samples of unknown depth.—A few samples of weathered core were collected in 1942 at the site of the boring. These consist of anhydrite with carbonates. It is not always possible to distinguish with certainty to which of the upper two evaporite beds such samples belong, so they have been largely omitted from this account.

Well cuttings taken at intervals of 5 to 10 feet, covering the whole depth range of the evaporites. These, of course, consist of insoluble residues.

My sincere thanks are due to Imperial Chemical Industries and to the D'Arcy Exploration Company for supplying material for examina-

tion and for allowing me to use data obtained by them. It is indeed fortunate that the D'Arcy Exploration Company, recognizing the possibility of the presence of potassium salts in this part of the salt field, undertook continuous coring throughout the salt series in their boring for oil. The analytical details of KCl percentages were provided by I.C.I., and are quoted with their kind permission. I am indebted to the Durham Colleges for a research grant, and to Mr. G. O'Neill for taking the photographs which illustrate this paper.

II. General succession in the middle evaporite bed.

The general succession is given in the following table. The relative distribution of anhydrite and halite in the lower part of the succession is based partly on data given in the D'Arcy Exploration Company's detailed log of the boring. The rest of the succession is based on a study of the available core samples and well cuttings

	Greenish-grey and red salt clay.	Thickness in feet.
3910 feet		
Upper halite zone	Brown, red, and colourless halite, with greenish-grey clay. (Many patches of blood-red halite, stained by haematite inclusions, in a specimen from 3916–18 feet)	8
3918 feet		
Halite-sylvine zone	Brown halite and sylvine, with greenish-grey clay	4
	No cores available	28
	Brown halite with subordinate sylvine and greenish-grey clay	14
3964 feet		
Halite zone	Brown and pink halite with patches of blood-red halite, the colour being due to haematite inclusions. Subordinate patches of anhydrite and clay, the anhydrite becoming small in amount towards the top. Very scarce sylvine near top ...	78
	Brown, gray, and colourless halite, with subordinate irregular layers and patches of anhydrite	53
	Alternating halite and anhydrite	3
	Brown halite with streaks of anhydrite	8
4106 feet		
Halite-anhydrite zone	Anhydrite and halite, with layers containing anhydrite-halite pseudomorphs after gypsum ...	30
	Colourless halite with streaks of anhydrite ...	4
	Anhydrite and halite	2
	Colourless halite with streaks of anhydrite ...	4
	Anhydrite and halite	7
	Colourless halite	2

4155 feet		
Anhydrite zone	Anhydrite with subordinate dolomite	41
4196 feet		
	Dolomite and dolomitic marl (Upper Magnesian Limestone).	

It can be seen that the succession is simpler, in general, than that of the lower evaporite bed (Stewart, 1949, p. 625), which is believed to represent the products of several interrupted evaporation cycles. A later stage in the evaporation series, however, has been reached—the formation of potassium chlorides. The absence of a polyhalite zone is striking. The polyhalite zone of the lower evaporite bed, before the formation of secondary polyhalite, was perhaps similar in character to the halite-anhydrite zone of the middle evaporites, with its pseudomorphs after gypsum. The absence of kieserite is also noteworthy, and, indeed, the relative absence of magnesium-bearing minerals other than carbonates. Carnallite is present only in small quantity in the halite-sylvine and halite zones, but, of course, magnesium salts may be present in larger quantity in the large part of the halite-sylvine zone from which no core samples are available (3922–50 feet). Since no special brine was used in the drilling, the very poor core recovery in this range would indicate the presence of highly soluble minerals.

The distribution of potassium.

The following details were provided by Imperial Chemical Industries, and are published with their kind permission. Each 6-inch specimen was cut in half longitudinally, and the material for analysis scraped from the flat surfaces so obtained:

	Depth below surface in feet.	Total KCl percentage.	KCl percentage on water-soluble material.
I.	3910	0·28	0·36
	3912	0·21	0·28
	3914	0·28	0·34
	3916	0·21	0·22
II.	3918	34·0	36·0
	3920	10·1	11·8
	3950	3·28	3·40
	3952	4·85	5·01
	3954	0·71	0·72
	3956	4·20	4·32
	3958	3·94	4·00
	3960	3·66	3·79
	3962	1·89	1·98
	3964	1·44	1·47

	Depth below surface in feet.	Total KCl percentage.	KCl percentage on water-soluble material.
III.	3970	0·63	0·64
	3972	1·96	1·99
	3974	0·41	0·42
	3976	0·46	0·47
	3978	0·38	0·40
	3980	0·60	0·61
	3982	0·70	0·70
	3984	0·70	0·70
	3986	1·28	1·34
	3988	0·40	0·41
	3990	0·41	0·42
	3992	0·41	0·42
	3994	0·40	0·41
	3996	0·41	0·42
	3998	0·30	0·31

	Depth below surface in feet.	Total KCl percentage.	KCl percentage on water-soluble material.
	4000	0·29	0·30
	4002	0·38	0·39
	4004	0·35	0·35
	4006	0·35	0·35
	4010	0·36	0·39
	4012	0·29	0·30
	4014	0·30	0·31
	4016	0·33	0·34
	4018	0·35	0·36
III.	4020	0·30	0·31
	4022	0·33	0·34
	4024	0·33	0·36
	4026	0·27	0·31
	4028	0·29	0·39
	4030	0·30	0·32
	4032	0·30	0·32
	4034	0·33	0·34
	4036	0·30	0·33
	4038	0·29	0·30
	4040	0·30	0·31
	4042	0·27	0·28
	4044	0·30	0·32
	4050	0·30	0·31
	4052	0·32	0·34
	4054	0·29	0·30
	4056	0·33	0·35
	4058	0·30	0·31
	4060	0·29	0·32
	4062	0·30	0·32
	4064	0·35	0·35
	4070	0·33	0·34
	4072	0·32	0·33
	4074	0·29	0·34
	4076	0·33	0·38
	4078	0·27	0·31
	4080	0·24	0·29
	4082	0·24	0·29
	4090	0·29	0·30
III.	4092	0·29	0·30
	4094	0·25	0·26
	4098	0·30	0·31
	4100	0·32	0·32
	4102	0·25	0·27
	4104	0·32	0·32
	4106	0·30	0·30
	4108	0·10	0·18
	4118	0·03	0·08
	4120	0·06	0·12
	4122	0·05	0·11
	4124	0·06	0·11
	4126	0·21	0·23
IV.	4128	0·22	0·23
	4130	0·03	0·12
	4132	0·25	0·25
	4134	0·17	0·17
	4138	0·19	0·20
	4140	0·10	0·16
	4142	0·14	0·15
	4144	0·14	0·14

I. Upper halite zone. II. Halite-sylvine zone. III. Halite zone. IV. Halite-anhydrite zone.

Three specimens from the salt clay between the middle and upper evaporite beds gave the following results:

Depth in feet.	Total KCl%.	KCl% in soluble material.
3870	0·28	1·31
3878	1·47	3·99
3898	1·28	5·95

These figures indicate the presence of potassium salts in these marls, but no samples of these were adequately preserved.

III. The minerals.

Anhydrite is generally grey or white in colour. Normally it forms fine-grained aggregates of interlocking grains (0·001 to 0·2 mm.), often showing a tendency to express rectangular habit, and sometimes elongated. Other types are large bluish platy grains up to 2 cm. across, fan-like

aggregates of fibrous crystals, and elongated or tabular crystals sometimes showing small dome and pyramid faces. Euhedral grains are very often developed when in contact with chlorides, when replacing dolomite, or when enclosed in clay. The optical properties are normal.

Boracite has been described by E. M. Guppy (1944) from well cuttings from the depth range 4090–4160 feet, with a chemical analysis by C. O. Harvey (see also Harvey, 1943). In the present investigation boracite has been found in small, but nearly constant, quantity in all samples of well cuttings from 3935–4160 feet. A few grains are also present in cuttings from the basal anhydrite zone. About three grams of the mineral have been isolated by hand-picking from the cuttings. It is a striking fact that no trace of the mineral has been found in the core samples, although a thorough search has been made.

The upper limit of a mineral's range can be determined fairly accurately by a study of well cuttings. However, if no special brine is used in the drilling, as was the case with this boring, there is likely to be considerable solution of the salts at the sides of the borehole as the drilling proceeds, and insoluble minerals will tend to be present in well cuttings from beneath the lower limit of their range. It will be noticed that the upper limit, 3935 feet, lies in the gap in the halite-sylvine zone from which no cores were recovered. It is therefore quite probable that the boracite occurs largely, if not entirely, somewhere within the range 3935–50 feet, in the halite-sylvine zone.

The boracite forms small isolated crystals and crystal groups, which are never attached to anhydrite or clay, and were presumably embedded in chlorides. The individual crystals are small (up to 1 mm.) and colourless, pale green, or red. They show pseudo-cubic forms resembling combinations of (100), (110), and (111). Only the very small crystals show nearly plane faces, the larger ones having irregular and curved faces sometimes made up of a large number of individuals in approximately parallel growth. Aggregates of crystals reach $\frac{1}{2}$ inch across. Determinations of optical properties on crushed grains agree with those given by Guppy: for sodium-light, α 1·658, β 1·662, γ 1·668, positive 2V 82°, within the limits of experimental error. The mineral shows complex twinning; often sector twinning with each sector polysynthetically twinned. Owing to overlapping of twin lamellae, only very thin fragments show extinction.

Carnallite occurs in small quantity from 3918 to 4022 feet. The mineral normally forms single grains, and more rarely aggregates of small irregular grains. It is normally light yellow in colour, occasionally

black. The largest grain of fresh carnallite seen is 5 mm. in length. The mineral may be detected readily by cutting a flat dry surface on the specimen and observing the development of rings of moisture round the small carnallite grains, due to their rapid deliquescence. The small grains can then be gouged out and determined by optical methods. Under the microscope its very low refractive indices, bright interference colours, and lack of cleavage will distinguish it from any other mineral yet found in these rocks. Many grains show simple twinning or polysynthetic twinning like that of the plagioclase felspars. Crushed fragments deliquesce rapidly and dissolve with the precipitation of small cubes of sylvine. Refractive index measurements gave the values α 1·466, β 1·472, γ 1·496 (for sodium-light). The average of five measurements on the universal stage gave positive 2V 72°.

Dolomite is the carbonate of the lower part of this evaporite bed, giving place to magnesite in the middle of the halite zone. The two carbonates have not been found together. The dolomite normally forms minute rounded granules (0·001 to 0·01 mm.). In the anhydrite zone much larger rhombohedra are also present. Twelve refractive-index measurements show no significant variation from the value ω 1·682.

Gypsum has not been found in this evaporite bed, but layers containing anhydrite-halite pseudomorphs after early gypsum occur in the upper part of the halite-anhydrite zone, from 4106–36 feet.

Haematite inclusions are present in much of the sylvine, some of the halite and boracite, and occasionally in quartz. Haematite occurs from 4042 feet to the top of the evaporite bed. Usually the mineral forms very thin hexagonal plates up to 0·3 mm. in diameter, but occasionally the crystals are rod-like. Mr. E. A. Vincent has kindly confirmed the identification by means of X-ray powder photographs. The mineral gives a uniaxial negative interference figure, and is pleochroic with ω dark brown, ϵ red.

Halite is restricted to interstitial traces in the anhydrite zone, but is by far the most abundant mineral of the rest of this evaporite bed. It forms grains up to several inches in size, normally with irregular margins. It is granular and colourless in the lower part of the halite-anhydrite zone, becoming faintly brown in tinge towards the top of this zone. In the higher zones it is most commonly light brown, but varies from colourless to grey, pink, brick-red, and blood-red, the last variety being full of haematite inclusions. Liquid-filled cavities, often cubic, are common, and are frequently arranged in zones parallel to the cleavages. Strain birefringence has not been observed.

Magnesite occurs from about 4036 feet in the halite zone to the top of the bed. The mineral forms crystals up to 0·5 mm. in size, which are usually combinations of hexagonal prism and base, markedly tabular parallel to (0001). Sometimes rhombohedral faces are developed. Eight refractive-index measurements give ω 1·700–1·702, indicating almost pure magnesite. Chemical tests show the virtual absence of calcium.

Magnetite. Tiny opaque grains have been found in a slice from 4140–42 feet, arranged radially round the edges of small pyrite grains enclosed in halite. Study of crushed material shows that these opaque particles are highly magnetic, and they are therefore provisionally identified as magnetite.

Pyrite. Small quantities of this mineral have been found in three rocks. Irregular grains, up to 0·2 mm. in size, occur enclosed in halite and lying at the contact of halite with anhydrite, in a slice from 4140–42 feet. The pyrite grains are surrounded by tiny radiating particles of magnetite. A single euhedral grain, 3 mm. across, is enclosed in halite in a specimen from 3986–88 feet. Several irregular grains, up to 1 mm. in size, have been found in a halite-sylvine-carnallite-rock at 3920–22 feet, enclosed in halite, and at the contact of halite, sylvine, and clay.

Quartz occurs from 4060 feet to the top of the evaporite bed, increasing in quantity towards the top. The mineral forms doubly-terminated prismatic crystals up to 2 mm. in length, showing the forms $(10\bar{1}0)$, $(10\bar{1}1)$, $(01\bar{1}1)$, $(11\bar{2}1)$, when enclosed in halite or sylvine.

Sylvine (recorded by Lees and Taitt, 1946) occurs in all specimens from the halite-sylvine zone, and in very small quantity in a few specimens near the top of the halite zone. The mineral forms irregular grains occasionally reaching $1\frac{1}{2}$ inches in size. Where free from haematite, it has a translucent milky-white colour, which has not been observed in halite. Much of it, however, is red, due to a multitude of tiny haematite inclusions. The red sylvine can be easily confused with red halite in hand-specimen, but the astringent taste is distinctive. It is readily determined optically by its isotropic character, perfect cubic cleavages, and low refractive index (1·490). It has a much higher surface relief than halite in thin sections. Liquid-filled cavities are not so frequently present as in halite.

IV. Rocks of the anhydrite zone.

Samples from 4160 and 4176–78 feet, and a third specimen from the lower part of the zone (4176–94 feet; exact depth unknown) were preserved. Several other specimens of unknown depth, collected at the

site of the boring, can be referred to this zone by comparison with the well cuttings.

The specimens and well cuttings show that subordinate dolomite is present throughout this zone, increasing in amount towards the base. The dolomite occurs as two distinct varieties: firstly, aggregates of minute rounded grains up to 0·01 mm. in diameter, coloured black in hand-specimen due to carbonaceous impurity; secondly, much larger clear euhedral rhombohedra, up to 1·5 mm. across, embedded in an anhydrite groundmass. A small quantity of interstitial halite is present in some cases.

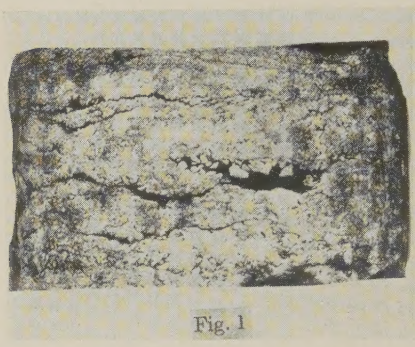

Fig. 1

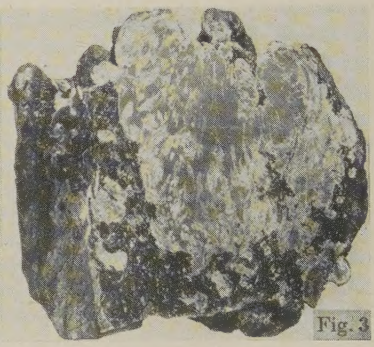

Fig. 3

Fig. 1. Anhydrite-dolomite-rock, showing layers of black dolomite partly replaced by anhydrite. Anhydrite zone, 4176–94 feet. $\times\frac{1}{2}$.

Fig. 3. Anhydrite-dolomite-rock with some interstitial halite (white and grey), broken up and partially replaced by halite (black). Ground and chalked surface. Halite-anhydrite zone, 4140–42 feet. $\times\frac{1}{2}$. (Upper surface at right-hand side.)

In the specimen from 4176–94 feet (figs. 1, 2A), fine-grained dolomite (about 20 % of the rock) forms nearly horizontal layers up to 5 mm. thick, alternating with layers composed largely of an interlocking mass of elongated anhydrite grains up to 1 mm. in length. The layering is believed to represent original bedding. In detail, however, the dolomite layers are irregular and patchy, and in slice it is evident that the dolomite has been partly replaced by anhydrite. The layers are broken up by patches and thin veins of anhydrite, and anhydrite crystals have developed inside the dolomite areas. Where the original dolomite was rich in carbonaceous impurity, the replacing anhydrite contains a fine dust of this material, in marked contrast to the clear anhydrite of the anhydrite layers. A few clear dolomite rhombohedra, up to 0·3 mm.

across, are enclosed in the clear anhydrite, but not in the fine-grained dolomite layers.

The rock from 4160 feet (fig. 2B) shows numerous rhombohedral porphyroblasts of dolomite, up to 1·5 mm. across, conspicuous in the hand-specimen. They are embedded in fine-grained anhydrite and enclose small anhydrite grains. There is relatively little residual fine-grained dolomite, broken up into irregular patches full of elongated laths of anhydrite

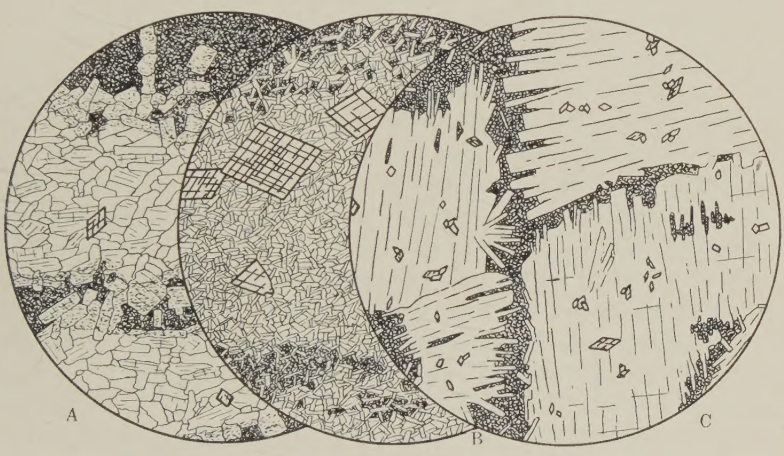

FIG. 2. Anhydrite-dolomite-rocks from the anhydrite zone. ×12.

A. Layers of fine-grained dolomite are partly replaced by anhydrite, the secondary anhydrite containing a dust of carbonaceous impurities. Larger euhedral dolomite grains are also present. 4176–94 feet.

B. Fine-grained dolomite is broken up by anhydrite, and dolomite porphyroblasts are enclosed in an anhydrite matrix. 4160 feet.

C. Coarse anhydrite replaces fine-grained dolomite and encloses euhedral dolomite. 4176–78 feet.

The rock from 4176–78 feet (fig. 2C) resembles the anhydrite-dolomite-rocks with coarse fibro-radiate anhydrite found near the base of the lower evaporite bed. The anhydrite forms large platy grains, with ragged edges, up to 2 cm. across, of slightly bluish tinge. In slice it can be seen that these grains consist of aggregates of fibrous crystals in approximately parallel growth. Many of the plates are broken down, or partially broken down, to masses of small irregular grains. There is about 20 % of dolomite. Between the large anhydrite masses are residual patches of fine-grained dolomite, and from the edges of the anhydrite masses fibrous and lath-

shaped anhydrite crystals push into the dolomite areas and break them up to form interstitial patches. Independent euhedral anhydrite laths develop inside the dolomite patches. Enclosed in the anhydrite, but not in the residual dolomite, are abundant euhedral dolomite rhombohedra up to 0·1 mm. across.

Some of the specimens of unknown depth are of this general type, showing coarse anhydrite, sometimes in radiate bunches, replacing fine-grained dolomite, with or without the development of euhedral dolomite porphyroblasts. Other specimens contain euhedral dolomite without any of the fine-grained variety, and some of these consist largely of platy anhydrite with rough horizontal platy parallelism.

The apparent bedding of the fine-grained dolomite and anhydrite would indicate that this type of dolomite is a primary precipitate. It seems probable that dolomite and anhydrite formed alternating primary layers, the dolomite being then partially replaced by anhydrite. The larger euhedral dolomite rhombohedra could be due to recrystallization of excess $CaMg(CO_3)_2$ attendant on the replacement (like the anhydrite porphyroblasts found in the polyhalite-bearing rocks of the lower evaporite bed; Stewart, 1949, p. 656).

An alternative view would be that the euhedral dolomite is due to recrystallization of the fine-grained dolomite without replacement. However, in the rock from 4176–94 feet some of the replacing anhydrite is dusted with impurities from the pre-existing dolomite, so that it is possible to make an approximate estimate of the minimum amount of dolomite replaced, and it is found that there is not nearly sufficient euhedral dolomite to account for recrystallization of replaced dolomite. Some dolomite must have been removed.

Whether the replacement was contemporaneous or later is a question which must be left open. It is possible that as each succeeding layer of anhydrite was deposited, the dolomite layer below was in part contemporaneously replaced by anhydrite, and some dolomite taken into solution. In the specimen from 4176–94 feet, more replacement seems to have taken place on one surface of each layer than on the other. If these surfaces are the upper surfaces of the succession, this would indicate contemporaneous replacement, but the evidence from this one 6-inch specimen is not sufficient to justify a definite conclusion.

V. Rocks of the halite-anhydrite zone.

This zone consists of a layered series of bedded anhydrite-dolomite-rocks; halite-rich rocks with subordinate anhydrite; halite-anhydrite-

dolomite-rocks with structures indicating movement of halite and replacement of anhydrite and dolomite by halite; and rocks with halite-anhydrite pseudomorphs after gypsum. Apart from the three main minerals, halite, anhydrite, and dolomite, there is only a very small quantity of pyrite and magnetite in some specimens, and a trace of detrital impurity. Fifteen specimens give an average of 0·14 % KCl.

(1) *Anhydrite-dolomite-rocks.*

Layers of compact anhydrite-dolomite-rock, up to a few inches thick, alternate with layers containing much halite. The anhydrite-dolomite-rocks generally show bedding structures due to alternation of layers rich and poor in dolomite. Most of these rocks are fine-grained. The anhydrite forms an interlocking mass of irregular grains (0·001 to 0·1 mm.), sometimes elongated. Tiny granules of dolomite (up to 0·01 mm.) are enclosed in anhydrite or lie between anhydrite grains, some layers containing up to 70 % of the mineral. In most cases, however, dolomite is subordinate in amount. A little interstitial halite is sometimes present, and there is dark dusty impurity in the dolomite-rich layers.

It is likely that the anhydrite and dolomite of these layers are primary precipitates, though some recrystallization of anhydrite may have taken place. In most slices there is no evidence of replacement of dolomite by anhydrite. In a few cases euhedral anhydrite crystals have developed in dolomite-rich areas and broken them up. There has been some replacement of anhydrite and dolomite by halite (pl. X, fig. 16).

(2) *Halite-rich rocks.*

Many layers consist largely of halite, without evidence of early gypsum. In the lower part of the zone, the halite is colourless and forms a granular mass of interlocking grains, up to 1 cm. across. In several specimens of this type the halite grains are flattened and show definite rough platy parallelism, imparting a rude horizontal schistose structure to the rock. There is also in some cases a trace of linear parallelism, the grains tending to lie with their longest dimensions in parallel orientation. This would imply recrystallization under pressure, or movement under the load of the overlying rocks. These structures are not found in the upper part of the zone, where halite of a brownish tinge is often coarsely crystalline, the grains reaching 3 inches in width.

Slices show that some anhydrite is nearly always present, usually as stumpy rectangular euhedral grains, often concentrated along intergranular boundaries of the halite.

(3) *Deformation and replacement of anhydrite-dolomite-rocks by halite.*

The halite-rich rocks of the lower part of the zone frequently enclose contorted and squeezed-out layers of anhydrite-dolomite-rock, and there are all gradations from halite-rich rock to anhydrite-dolomite-rock. Thin anhydrite-dolomite layers are often folded and crenulated, and those enclosed in schistose granular halite are sometimes broken up

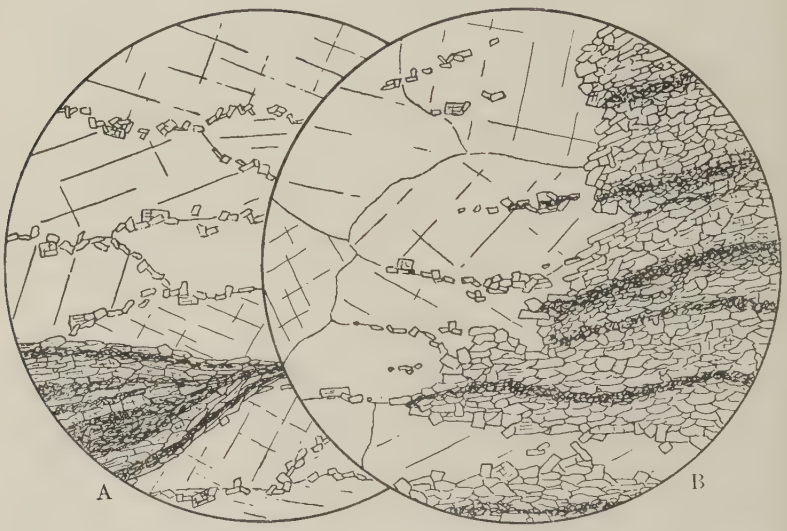

Fig. 4. A. Halite-anhydrite-rock with deformed lenticle of anhydrite-dolomite-rock. The halite grains are markedly flattened. Halite-anhydrite zone, 4142–44 feet. ×14.

B. Anhydrite-dolomite-rock partially replaced by halite, showing recrystallized anhydrite relics. Halite-anhydrite zone, 4140–42 feet. ×14.

into elongated patches and lenticles arranged parallel to the schistosity of the halite. Slices show that the bedding structures in the anhydrite-dolomite relics have been deformed so that they are parallel to the edges of the lenticles (fig. 4A), and the anhydrite of the thinner layers has developed a platy parallelism of the grains. The anhydrite tends to become coarser and euhedral when in contact with halite, indicating late recrystallization.

The invasion of anhydrite-dolomite-rock by halite is not only mechanical (figs. 3, 4B). Some layers of anhydrite-dolomite-rock come to an abrupt end, cut off by granular halite. The bedding structures are cut off sharply, and not curved round parallel to the edges of the layers.

Sometimes thin relict streaks of anhydrite grains continue from the anhydrite-dolomite layers into the adjacent halite, parallel to the original bedding. Thin veinlets of halite penetrate the anhydrite-dolomite-rock, chiefly following the bedding, but sometimes cutting across it. It appears that halite has replaced both anhydrite and dolomite.

Some of the rocks showing replacement have suffered subsequent deformation, the halite developing a platy or linear parallelism and the bedding of the anhydrite-dolomite-rock becoming distorted. This does not date the replacement as it is possible that deformation under cover may have taken place at various times, either simultaneously with, or both before and after replacement. In one specimen from the upper part of the zone, containing anhydrite-halite pseudomorphs after gypsum, a vein of granular halite, similar in type to that of the rocks described here, cuts through the coarse secondary halite of the pseudomorphs and the anhydrite-dolomite matrix. This indicates that the replacements involving granular halite are later than the development of secondary halite in the rocks with gypsum pseudomorphs.

The relative restriction of evidence of movement of halite to the lower part of this zone is striking. Perhaps the basal anhydrite zone provided a solid and unyielding block against which the halite gave way to pressure by flow, while the upper zones were able to adjust themselves by recrystallization without significant flow.

(4) *Rocks containing pseudomorphs after gypsum.*

Layers containing anhydrite-halite pseudomorphs, which are believed to represent early gypsum crystals, are found in the upper 30 feet of this zone. These layers vary from about an inch to five inches in thickness, and alternate with layers of anhydrite-dolomite-rock and halite-rich rock.

The pseudomorphs are similar in shape to those of the polyhalite zone of the lower evaporite bed, and are very like those described by Schaller and Henderson (1932) from the Texas–New Mexico potash-field. They range from less than an inch to $3\frac{1}{2}$ inches in length, and in some cases reach $\frac{3}{4}$ inch in width. In most cases they are arranged in an approximately vertical position in the core, although in one specimen pseudomorphs up to 3 inches in length form a radiating fan-shaped group (pl. X, fig. 13). They are irregularly six-sided in cross-section (pl. X, fig. 15) and often show terminations like those of swallow-tail twins of gypsum (fig. 5). A few of these pseudomorphs consist entirely of anhydrite; some have a centre of halite and an outer zone of anhydrite; but most show a well-

marked herring-bone structure, consisting of alternate ribs of anhydrite and halite arranged parallel to the swallow-tail twin terminal faces, often with a central vertical partition of anhydrite along the composition-plane of the twin. The relative amounts of halite and anhydrite vary considerably in different pseudomorphs and in different parts of the same pseudomorph, and in many cases the anhydrite ribs are very thin, or broken and discontinuous (pl. X, fig. 14).

Fig. 5

Fig. 6

Fig. 5. Halite-anhydrite-rock with pseudomorphs after swallow-tail twins of gypsum. The surface of the core has been etched by the drilling water, the halite (dark) being partially dissolved. Halite-anhydrite zone, 4108–10 feet. $\times\frac{1}{2}$.

Fig. 6. Alternating crumpled layers of halite-anhydrite-rock and anhydrite-dolomite-rock. Etched surface of core. Halite zone, 4096–98 feet. $\times\frac{1}{2}$. (Upper surface at right-hand side.)

In most cases the interstices between the pseudomorphs consist of anhydrite-dolomite-rock, which occasionally shows irregular bedding structure upturned at the edges of the pseudomorphs. It is believed that in such rocks the gypsum crystals grew at the expense of anhydrite, in anhydrite-dolomite-rock, when the latter was in a mushy state, the replacement being an early one (see Schaller and Henderson, 1932, p. 34; Stewart, 1949, p. 653). It is possible that the ribs of anhydrite (which contain some dolomite) in the herring-bone structures represent relics of the original anhydrite-dolomite-rock. Dunham (1948, p. 225, and plate X, fig. 1) has described peculiar zoned porphyroblasts of

gypsum containing zones of anhydrite relics in anhydrite-dolomite-rock from the outer part of this salt-field.

The gypsum has been subsequently converted to halite. In slice, it can be seen that this halite encloses euhedral anhydrite crystals, and that while the interiors of the anhydrite ribs consist of fine-grained aggregates of irregular grains, the edges consist of elongated euhedral crystals, up to 0·5 mm. long, projecting into the adjacent halite. It seems likely that recrystallization of anhydrite has accompanied the appearance of secondary halite. The recrystallized anhydrite has cleared itself of impurity, and dolomite and other impurities are concentrated in the centres of the anhydrite ribs (pl. X, fig. 14). Several adjacent pseudomorphs often contain coarsely crystalline halite in common orientation. This feature was noted by Schaller and Henderson (1932, p. 34), and interpreted as indicating that the pseudomorphs may be infiltration pseudomorphs, the gypsum being dissolved, leaving cavities which were later filled with halite. However, they also found evidence indicating a more intimate type of replacement.

In the rocks described above, we have the replacement sequence: anhydrite→gypsum→halite. In other layers, however, the interstices between the pseudomorphs consist of halite, and some of the pseudomorphs consist of anhydrite, others of halite with an outer zone of anhydrite. In these cases it appears likely that the gypsum was a primary mineral, the replacements being: gypsum→anhydrite and gypsum→anhydrite + halite.

The pseudomorphous layers show relatively little evidence of later distortion. Some of the pseudomorphs have been deformed, but there appears to have been little movement in the upper part of the zone.

(5) *General considerations on zones containing evidence of early gypsum.*

The upper part of the halite-anhydrite zone appears to be similar to part of the polyhalite zone of the lower evaporite bed, except that the former contains no polyhalite. A similar zone has been found in the upper evaporite bed. In all three beds, these zones lie in similar positions, above the basal anhydrite and below the chloride-rich zones.

We do not know which were the upper, and which the lower, surfaces of the specimens from the Eskdale boring, so that we cannot give the detailed order of succession of the layers. However, in the Texas–New Mexico potash-field Schaller and Henderson (1932, p. 36) show that in the typical succession a basal layer of clay is succeeded upwards by anhydrite, which is followed by anhydrite with halite pseudomorphs

after gypsum, and then by halite. Polyhalite may or may not be present. Evidence of primary gypsum was not found, but it was considered that the gypsum had formed at an early stage in unconsolidated anhydrite mud. The restriction of gypsum to the upper parts of the basal anhydrite layers led the authors to suggest that the gypsum may have been derived from earlier polyhalite.

The presence of these zones of early gypsum in similar positions in all three evaporite beds at Eskdale, and of similar rocks in the Texas New Mexico deposits, suggests that a zone containing early gypsum may prove to be a normal stage in the formation of many marine evaporites: a stage at which slight rhythmic changes in conditions may cause alternate production of anhydrite-dolomite and halite-gypsum associations.

The following tentative suggestions indicate how the observed rocks may have formed. Take the ideal depositional succession to be: halite—halite-gypsum—anhydrite-carbonate. After precipitation of a basal layer of anhydrite-carbonate-rock (underlain by clay in the Texas–New Mexico case) the changing composition of the brines or some other factor caused halite-gypsum-rock to be deposited. At the same time the anhydrite already formed became unstable in the gypsum-depositing brines, and gypsum crystals grew at the expense of this anhydrite in the unconsolidated upper part of the basal anhydrite-carbonate layer. As the overlying brine changed further in composition, it ceased to deposit gypsum, and formed a layer of halite. The gypsum of the earlier layers, in turn, became unstable and was replaced by halite and anhydrite. A reversal of conditions would cause deposition of anhydrite and carbonate again and a repetition of the sequence. By periodic changes in conditions, a rhythmically layered series of units showing the following sequence would be built up: halite-rock halite-rock with halite-anhydrite pseudomorphs after gypsum anhydrite-carbonate-rock with halite-anhydrite pseudomorphs after gypsum—anhydrite-carbonate-rock.

The absence of any particular layer could be explained by postulating re-solution by the overlying brines. The restriction of secondary gypsum to the upper part of the basal anhydrite in the Texas–New Mexico succession could be explained by assuming that the brines responsible for its development percolated from above, and only reached a certain distance into this anhydrite layer. There would be no need to derive the secondary gypsum from earlier polyhalite.

Once again the observed succession does not agree with that expected

on experimental grounds. Posnjak (1938, 1940) has shown that gypsum would be expected to form before anhydrite on evaporation of sea-water. We should therefore expect a basal layer of gypsum-carbonate-rock, and anhydrite, if present, should occur above this and in the halite-bearing rocks. There is, however, no evidence that the anhydrite of the bedded anhydrite-carbonate-rocks has been derived from gypsum.

VI. Rocks of the halite zone.

The lower limit of this thick zone has been fixed at the level where coarse brown halite, with only subordinate anhydrite-carbonate layers, appears, above the highest specimen containing pseudomorphs after gypsum (4108–10). The upper limit is where sylvine first occurs in significant amount. In 59 specimens the average KCl content is 0·39 %.

The minerals of the zone are halite, anhydrite, dolomite, magnesite, carnallite, sylvine, quartz, haematite, and pyrite. Greenish clay forms patches in the upper part of the zone.

In the lower part of the zone thin anhydrite-dolomite layers alternate with layers consisting largely of halite. The anhydrite-dolomite layers are similar to those of the zone below, the dolomite forming the usual minute granules. They are frequently contorted and occasionally faulted (figs. 6, 7). The halite-rich layers contain some euhedral anhydrite, but are largely composed of irregular grains of brown, pink, or colourless halite, often varying in size from a few millimetres to several inches in the same specimen. In other specimens the halite is even-grained.

The anhydrite-dolomite layers of the lower part of the zone become much scarcer upwards. They also become more highly contorted, and broken up, and are eventually represented by isolated interstitial patches between halite grains. Throughout the upper part of the zone halite forms over 80 % of the rocks, and the other minerals (except the chlorides), with detrital impurities, are largely confined to these interstitial patches. The halite tends to develop crystal boundaries against these aggregates, although the individual grains of the other minerals are euhedral towards halite. It would appear that recrystallization of halite has pushed the impurities into the interstices (figs. 8, 9A, 10A).

The upper part of the zone is mineralogically more complex than the lower. Halite becomes very coarse in places. Anhydrite varies from aggregates of irregular grains to spherulitic groups of elongated crystals, and platy crystals. Magnesite is the carbonate (dolomite being absent), and has been detected in most slices above 4036 feet. It is normally

present in very small quantity, but larger amounts are associated with anhydritic clay towards the top of the zone. It is mainly found in the interstitial areas, but some isolated crystals and crystal groups are enclosed in halite. It forms small tabular crystals or rhombohedra, up to 0·2 mm. diameter, and is euhedral towards all the other minerals. Quartz has been found in most slices above 4060 feet, and is present

Fig. 7

Fig. 8

Fig. 7. Anhydrite layers faulted and contorted due to recrystallization of adjacent halite. Etched surface of core. Halite zone, 4094–96 feet. $\times \frac{1}{2}$.

Fig. 8. Coarse halite with interstitial patches of anhydrite and clay. Polished surface. Halite zone, 4034–36 feet. $\times \frac{2}{3}$.

only in small quantity in the lower part of its range, but forms about 1 % of some slices in the upper part. It forms excellent doubly terminated crystals up to 0·5 mm. in length, when enclosed in halite. It is especially concentrated in, and at the edges of, interstitial areas, where crystals moulded on anhydrite and magnesite, towards which it is anhedral, project into halite (figs. 9B, 10A). This suggests late crystallization. Some of the silica may have been derived by solution of the detrital material with which it is so often associated.

Carnallite is present in most specimens from 4022 feet upwards, in

very small quantity, as small rounded or irregular grains or aggregates enclosed in halite. It has not been found in contact with any other mineral. Sylvine has been observed in a few specimens near the top of the zone, where it forms grains up to about 5 mm. across, showing irregular boundaries against halite.

Fig. 9. Rocks from the halite zone. ×14.

A. Halite-anhydrite-rock with carbonaceous impurities. The halite shows crystal boundaries against anhydrite aggregates, although individual anhydrite crystals are euhedral towards halite. Veinlets of halite traverse the anhydrite area. 4078–80 feet.

B. Halite-anhydrite-rock with dark clay, and a dust of haematite inclusions in parts of the halite. A few euhedral quartz crystals project from the clay into halite. The arrangement of haematite inclusions suggests either that the large halite grain has formed by recrystallization of a number of smaller grains, or that halite has replaced smaller grains of some other mineral. 3978–80 feet.

The sylvine tends to have red margins, due to haematite inclusions, but most of the haematite of this zone is enclosed in halite, colouring it blood-red (fig. 9B). The haematite forms very thin hexagonal plates and rods, usually extremely small, but occasionally reaching 0·1 mm. across. At 4042 feet, small patches of blood-red halite, crowded with countless inclusions, make an appearance, and these get more numerous towards the top of the zone, reaching about a centimetre in size. The

inclusions are sometimes arranged in zones parallel to the halite cleavages, but this zoning is commonly distorted, and all stages are found between cubic zoning and irregular wavy bands of haematite. The mineral is often concentrated round interstitial patches of greenish-grey clay, and is enclosed in any halite present within these patches. Perhaps iron oxide has been leached out of the clay and precipitated as haematite with the neighbouring halite (Schaller and Henderson, 1932, p. 24). On the other hand, clay is not always present in the haematite-rich patches. Some of the quartz enclosed in red halite encloses haematite, zoned parallel to its crystal faces.

Recrystallization and replacement.

The above details show that there has been large-scale recrystallization of halite in this zone. The evidence includes contortion and faulting of anhydrite-carbonate layers in the lower part of the zone; breaking up of the layers to form interstitial patches, often showing streaking out of impurities, between halite grains throughout the upper part of the zone; ejection of detrital material with magnesite, &c., from halite into the interstices; distortion and disruption of the zoning of haematite inclusions. It is believed that recrystallization of halite has taken place throughout the zone. None of the halite shows linear or platy parallelism of the grains.

The faulting of anhydrite layers shows that recrystallization took place when the anhydrite was sufficiently solid to yield by fracture. That recrystallization took place at various times is shown by the occurrence of aggregates of clay, anhydrite, and other minerals, having the angular shape of interstitial patches, completely enclosed in large grains of halite formed by later recrystallization (fig. 10B). A similar thing is seen in the case of haematite inclusions, where the later light-coloured halite has failed to re-orientate the inclusions in the earlier angular red patches enclosed in it.

Some recrystallization of anhydrite has taken place, giving rise to euhedral crystals cutting a finer-grained anhydrite matrix, and euhedral crystals projecting into halite from the edges of anhydrite layers and patches. In one slice (pl. X, fig. 17) interstitial aggregates of small elongated anhydrite grains show a honeycomb network of detrital impurities, suggesting simultaneous recrystallization of anhydrite from a number of centres, the impurities being pushed out and concentrated between the areas of recrystallization.

Some replacement of anhydrite by halite is indicated by irregular veinlets of halite cutting anhydrite areas. In a few cases anhydrite grains have been found to be broken up into several pieces, with common optical orientation, by invading halite.

The haematite of other salt deposits tends to be concentrated particularly in the magnesium and potassium minerals, and in the Eskdale

Fig. 10. Rocks from the halite zone. ×14.

A. Halite-anhydrite-magnesite-quartz-rock, showing large grains of halite, and an angular patch containing small elongated rectangular anhydrite crystals, magnesite plates (heavy boundaries), a few quartz crystals, and some detrital material. 3992–94 feet.

B. Halite-anhydrite-carnallite-rock with scarce magnesite. A carnallite grain (stippled) is enclosed in halite. The anhydrite is aggregated in angular patches enclosed in a large halite grain, indicating recrystallization of halite. 4016–18 feet.

evaporites it is found in sylvine and boracite, as well as in halite. It is perhaps possible that the haematite of this halite zone has been derived from some pre-existing magnesium- or potassium-bearing salt which has been replaced by halite (fig. 9B is suggestive). No residual patches of such a mineral have been found in the halite zone, but the widespread recrystallization of halite would tend to destroy most of the evidence of early replacements in this zone.

VII. Rocks of the halite-sylvine zone.

Ten samples from this zone are available. There are two sylvine-rich samples from near the top of the zone (3918–20 feet, 34 % KCl; 3920–22 feet, 10·1 % KCl). Below this level there is a gap of 28 feet from which there are no available specimens. This is followed by 14 feet represented by eight specimens with an average KCl content of 3 % (maximum 4·85 % KCl, minimum 0·71 % KCl). The poor core recovery may well be due to the presence of highly soluble potassium and magnesium chlorides, leading to solution and disintegration of the core. The well cuttings give no information on the soluble minerals, so that we do not know what lies between 3922 and 3950 feet. They show, however, that boracite (not found in the core samples) appears at 3935 feet.

Most of these rocks show no layered structure, and consist largely of a coarse-grained mixture of halite and sylvine, showing evidence of widespread recrystallization. In such rocks the small amount of greenish-grey clay present is concentrated in interstitial patches, together with the accessory minerals other than carnallite and haematite. There are, however, some layers rich in clay, and there the chlorides tend to occur in smaller grains, about $\frac{1}{4}$ inch or less, with a tendency to aggregate in lenticular patches, elongated horizontally. This imparts a rough lamination to the rock.

The minerals of this zone are halite, sylvine, carnallite, anhydrite, magnesite, quartz, haematite, and pyrite. All these minerals may occur in the same rock. Apart from clay, halite and sylvine form over 90 % of all the specimens. Sylvine is most abundant in the uppermost specimen (3918–20 feet), where a three-inch layer consists of halite and sylvine in nearly equal proportions, with about 1 % of carnallite and only traces of magnesite and anhydrite.

The halite varies from colourless to reddish-brown, and the grains sometimes reach 2 inches in size. The sylvine is either translucent milky-white, or bright-red due to haematite inclusions. One grain of sylvine reaches $1\frac{1}{2}$ inches in size (3958–60 feet), but most are much smaller. Commonly the sylvine encloses tiny rounded inclusions of halite. The two minerals normally show interlocking irregular boundaries. In one slice of the specimen from 3918–20 feet they are intergrown in common orientation, the cleavages passing continuously through both minerals (fig. 11A). It is interesting that in this case the sylvine contains haematite, but the halite is clear. In several slices and hand-specimens a replacement relationship is seen, where sylvine grains have been broken

up into residual patches with common orientation (fig. 12A). Colourless halite often replaces sylvine with haematite inclusions, showing that the solutions responsible for the replacement were capable of removing haematite as well as sylvine (fig. 12C). More rarely the haematite inclusions persist in the secondary halite, occasionally in sharp-edged areas marking the boundaries of the original sylvine grains (fig. 12B).

FIG. 11. Halite-sylvine-carnallite-rocks from the halite-sylvine zone. 3918-20 feet.

A. Sylvine (lightly stippled) is intergrown with halite in common orientation, the sylvine containing plates and rods of haematite, while the halite is clear of inclusions. A small grain of carnallite (heavily stippled) is enclosed in halite. ×14.

B. A large carnallite grain lies between halite and sylvine, and several smaller carnallite grains are enclosed in halite. The shape of the carnallite grains suggests corrosion, but there is no definite evidence of replacement since the grains have different optical orientations. ×7.

Late veinlets and patches of halite in these rocks are normally colourless. Halite and sylvine may enclose any of the other minerals, showing irregular boundaries towards carnallite and pyrite, and being anhedral towards anhydrite, magnesite, and quartz.

Carnallite occurs only in small quantity (1 % or less), but is present in all specimens. Small single grains of the mineral, with irregular or rounded margins, are enclosed in halite, and more rarely sylvine, or lie between halite and sylvine grains (fig. 11). No definite evidence of a

replacement relationship between carnallite and the chlorides has been observed, and the mineral has not been found in contact with anhydrite, magnesite, or quartz.

Anhydrite is very scarce in the upper two rocks, only a few rectangular crystals being found. In the lower part of the zone it is more plentiful, but not abundant. In the lowest rocks, interstitial aggregates of small grains, like those of the halite zone, are found, but these give place higher up to radiating fan-like groups of elongated crystals up to 0·5 mm. in length. Sometimes these groups are aggregated together in shapes suggesting that they have been derived by recrystallization of interstitial patches. There is some evidence of replacement of anhydrite by halite, crystals of anhydrite being broken up into residual corroded relics (fig. 12A).

Magnesite is fairly abundant in the layers rich in clay (e.g. 3920–22 feet), but only a few small grains are found elsewhere. It forms thin hexagonal plates, tabular parallel to (0001), like the magnesite of the lower evaporite bed. These plates are rod-like in cross-section, and occasionally reach 0·5 mm. in diameter. Magnesite is euhedral towards all other minerals.

Quartz is especially associated with clay, and is abundant in layers rich in clay (fig. 12). It forms well-shaped crystals in contact with halite and sylvine, and is anhedral towards anhydrite and magnesite. Its special association with clay would suggest that the silica has been at least partly derived by leaching of the clay.

The distribution of haematite and the origin of sylvine.

Plates and rods of haematite, generally minute, but occasionally reaching 0·3 mm. in size, are found in most of the sylvine, and very little of the associated halite. The inclusions are usually unorientated, but are occasionally arranged in zones, often distorted, inclined to the cleavages of their host. Some small sylvine grains are crowded with minute inclusions, uniformly distributed, but most often the inclusions are concentrated near the borders of the grains, the centres being clear, and in these cases the inclusions are usually larger and fewer in number. Sometimes they are concentrated near the centres as well as at the edges of sylvine grains (pl. X, fig. 19), and they are frequently concentrated round foreign inclusions such as quartz, magnesite, and anhydrite grains (fig. 12C). They are not, however, enclosed in these minerals, except rarely in quartz. Occasionally a sylvine grain, with an outer zone of relatively large haematite flakes, contains a sharply

bounded central area crowded with minute haematite inclusions arranged in criss-cross sets of zones inclined to the cleavages (fig. 12c and pl. X, fig. 18). This suggests either that there are two generations of sylvine, or that the central area of sylvine has been derived from some other

Fig. 12. Rocks from the halite-sylvine zone. ×12.

A. Halite-sylvine-anhydrite-rock, showing three corroded relics of sylvine (stippled) in halite, and replacement of anhydrite (top left) by halite. 3962–64 feet.

B. Halite-sylvine-rock with magnesite, quartz, and clay, showing replacement of sylvine by halite. A zone of haematite inclusions marks the original boundary of the sylvine grain. 3918–20 feet.

C. Halite-sylvine-rock, with magnesite, quartz, anhydrite, haematite, and a considerable amount of interstitial clay. The edges of sylvine grains are crowded with haematite inclusions, which are also concentrated round foreign crystals included in sylvine. To the right is a sylvine grain with haematite-rich edges and a sharply bounded central area rich in zoned haematite inclusions, the zones being inclined to the sylvine cleavages. This suggests replacement of some pre-existing mineral by sylvine. Halite is practically free from haematite, and late veinlets and areas of halite have replaced sylvine and removed haematite. Magnesite plates (thick boundaries) and quartz (thin boundaries) are fairly abundant, and a few rectangular crystals of anhydrite are present. 3920–22 feet.

mineral. There is sometimes a tendency for haematite to be concentrated in interstitial areas like those of the halite zone, but this is not marked. Sylvine and halite tend to be more deeply coloured in rocks rich in clay than in other rocks, but even in layers containing practically no clay the sylvine normally contains some haematite.

In the Texas–New Mexico field haematite is stated to be largely confined to the potassium-bearing minerals (Schaller and Henderson, 1932; Mansfield, 1930; Phillips, 1947), being present in polyhalite, carnallite, leonite, langbeinite, and sylvine. It is also present in glauberite and to a very small degree in halite. Schaller and Henderson suggest that it may have a bacteriological origin, and also suggest that there are two generations of sylvine, the red variety replacing the colourless along intergranular boundaries, cracks, and cleavages. They never found the red variety surrounded by the colourless.

Johnsen (1909) ascribes orientated haematite inclusions in the Stassfurt carnallite to the oxidation of carnallite which contained some $FeCl_2$ as an isomorphous replacement of $MgCl_2$. Leonhardt and Tiemeyer (1938) record sylvine from Wathlingen, Hanover, in which the (0001) faces of haematite lie on the (100), (111), and (110) planes of sylvine. They conclude that the preferred orientation of the haematite shows that the mineral crystallized simultaneously with sylvine, and that therefore the sylvine and haematite were not derived from earlier carnallite.

In the Eskdale evaporites the concentration of haematite at both centres and margins of sylvine grains, and round foreign inclusions, would suggest recrystallization of sylvine with partial ejection of haematite, and not replacement of colourless by red sylvine. Secondary clear halite has replaced sylvine, but it is most unlikely that all the clear halite of this zone is secondary, and if sylvine is a primary mineral it must almost certainly have been deposited together with halite. In this case there would appear to be no reason why haematite should not be present in both minerals, unless the iron preferred to go into solid solution only in sylvine, and later separated as haematite. However, it seems unlikely that any appreciable amount of iron could be contained in solid solution in either sylvine or halite.

This raises the suggestion that the red sylvine found with colourless halite may have been formed by replacement of some pre-existing magnesium salt, such as carnallite. If Fe-Mg-carnallite were replaced by sylvine under oxidizing conditions, the iron oxide might well be precipitated as haematite on replacement, and the haematite might sometimes form orientated inclusions in the secondary sylvine. All of the red potassium salts of the Texas–New Mexico field, except sylvine, contain magnesium. Iron could perhaps replace magnesium in these salts, and calcium in glauberite, and separate later by oxidation and ex-solution as haematite.

It is, of course, not claimed that haematite cannot be deposited directly from salt solutions under certain circumstances, but it is suggested that red sylvine associated with other colourless primary salts such as halite may well be secondary. However, more decisive evidence is needed in the present case.

VIII. Rocks of the upper halite zone.

The four available specimens from this 8-foot zone have an average KCl content of only 0·25 %. They consist largely of halite, with a considerable quantity of clay. Magnesite and quartz are more abundant than in the lower zones, quartz forming about 3 % of some slices (pl. X, fig. 21). Anhydrite is scarce. Sylvine and carnallite are absent. The minerals have the usual relations to one another. Some anhydrite grains are partially replaced by halite, but no other direct evidence of replacement has been found.

The halite is relatively fine-grained (about 1 mm. to 1 cm.). Quartz crystals reach 2 mm. in length. Magnesite crystals are combinations of short hexagonal prisms with rhombohedra and base, up to 0·3 mm. in width. The halite is brick-red in colour in the upper three specimens. The lower specimen (3916–18) lies 2 feet above the specimen with 34 % KCl in the halite-sylvine zone. It contains only 0·21 % KCl, and consists of light brown halite (about 50 %), angular patches of blood-red halite full of haematite inclusions (about 45 %), and small quantities of quartz, magnesite, and anhydrite (about 5 %). The angular aggregation of haematite inclusions and distortion of zones of inclusions indicate recrystallization of halite (pl. X, fig. 20).

It seems likely that the haematite has in this case been derived from earlier sylvine or some earlier potassium or magnesium salt, although no relics of such a mineral have been found. It would be expected that with the incoming of the clastic material which sealed off the top of the evaporite bed, or with later percolation of ground-water through the clastic cover, there would be leaching of the more soluble salts in the upper part of the evaporite bed. This upper halite zone is believed to be such a 'cap' zone. Kainite, which is commonly formed by reaction of sylvine or carnallite with kieserite in the 'cap' zones of the German deposits, has not been found here, but the absence of kieserite would account for this.

IX. Summary and conclusions.

The evaporite bed is 286 feet thick, and is underlain by dolomite and overlain by salt clay. It has been divided into the following five zones: Upper halite zone, 8 feet; Halite-sylvine zone, 46 feet; Halite zone, 142 feet; Halite-anhydrite zone, 49 feet; Anhydrite zone, 41 feet.

The following minerals have been found: anhydrite, boracite, carnallite, dolomite, haematite, halite, magnesite, magnetite, pyrite, quartz, and sylvine. The major constituents are halite, anhydrite, carbonates, and sylvine. Quartz is unusually abundant. Dolomite is restricted to the lower part of the succession, and magnesite, quartz, carnallite, sylvine, and haematite to the upper part. Some of the silica and iron oxide necessary to the formation of quartz and haematite may have been derived by solution of detrital material.

Petrographic descriptions of the rocks are given. The following replacements have been recorded: Dolomite→anhydrite; Dolomite→halite; Anhydrite→halite; Anhydrite→gypsum→halite; Gypsum→anhydrite±halite; Sylvine→halite.

Replacement of dolomite by anhydrite has taken place in the anhydrite zone, and to a small extent in the halite-anhydrite zone. The formation of dolomite porphyroblasts has apparently accompanied the replacement. There is no definite evidence to date this replacement.

Replacement of anhydrite-dolomite-rock by halite has taken place in the halite-anhydrite zone, and some of the secondary halite has suffered subsequent deformation. This replacement is probably later than those involving gypsum.

Halite has replaced some anhydrite in all the upper four zones. Some of this replacement appears to have taken place during or after recrystallization of halite, and after earlier recrystallization of anhydrite. It seems that halite has replaced anhydrite at several different times.

The replacements involving gypsum took place in the upper part of the halite-anhydrite zone. It is considered that some of the gypsum was primary, and some derived from anhydrite in unconsolidated anhydrite-dolomite-rock, all of it being later converted to halite and anhydrite, probably contemporaneously with the formation of this zone. There is no evidence of early gypsum in the anhydrite zone.

The presence of zones containing early gypsum in all three Eskdale evaporite beds, and of similar rocks in the Texas–New Mexico potash-field, suggests that such a zone may prove to be a normal stage in the formation of many marine evaporites; a stage during which slight

rhythmic changes in conditions may cause alternate deposition of halite-, gypsum-halite-, and anhydrite-dolomite-rocks, with metasomatism of each earlier layer during the deposition of the one above it.

Replacement of sylvine by halite has been found in the halite-sylvine zone, and some of this appears to have taken place after recrystallization of halite and sylvine. Whether sylvine was primary or not cannot be decided. There is no direct evidence that it has replaced any pre-existing mineral, but the widespread recrystallization in the chloride zones would tend to destroy evidence of earlier replacements. The distribution of haematite suggests that the sylvine has been derived from some pre-existing magnesium-bearing salt, such as carnallite, containing ferrous chloride in solid solution, the latter separating as haematite during the replacement.

In the lower part of the halite-anhydrite zone, directional and deformational structures indicate recrystallization and 'flow' of halite under pressure, above the solid and unyielding anhydrite zone. There is evidence of wholesale recrystallization of halite, without directional structures, in the greater part of the chloride zones. Much recrystallization of anhydrite has taken place.

These evaporites provide the first record of significant quantities of potassium-bearing chlorides in Britain, and of the occurrence of sylvine, carnallite, and boracite.

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Explanation of Plates IX and X.

Photograph and photomicrographs of evaporites from the middle bed in the Eskdale no. 2 boring, east Yorkshire.

Plate IX.

Fig. 13. Anhydrite-dolomite-halite-rock with radiating group of distorted pseudomorphs after gypsum. The pseudomorphs show irregular 'herring-bone' structure, and consist of halite (black), and anhydrite and dolomite (white and grey). The interstices consist of anhydrite and dolomite. A late vein of halite is present near the base of the specimen. The surface has been ground with coarse carborundum, the halite polished with a wet cloth, and the specimen rubbed with chalk, which adheres to anhydrite and dolomite but not to halite. Halite-anhydrite zone, 4108–10 feet. Natural size.

Fig. 14. Part of large 'herring-bone' pseudomorph after gypsum. Irregular ribs of anhydrite (grey) and dolomite (black) alternate with halite areas. Halite-anhydrite zone, 4108–10 feet. ×10.

Fig. 15. Anhydrite-dolomite-halite-rock, showing cross-section of pseudomorph after gypsum. The gypsum crystals probably grew at the expense of anhydrite in anhydrite-dolomite-rock, pushing the dolomite (black) into the interstices. The gypsum was later replaced by anhydrite (grey) and halite (white). Halite-anhydrite zone, 4124–26 feet. ×10.

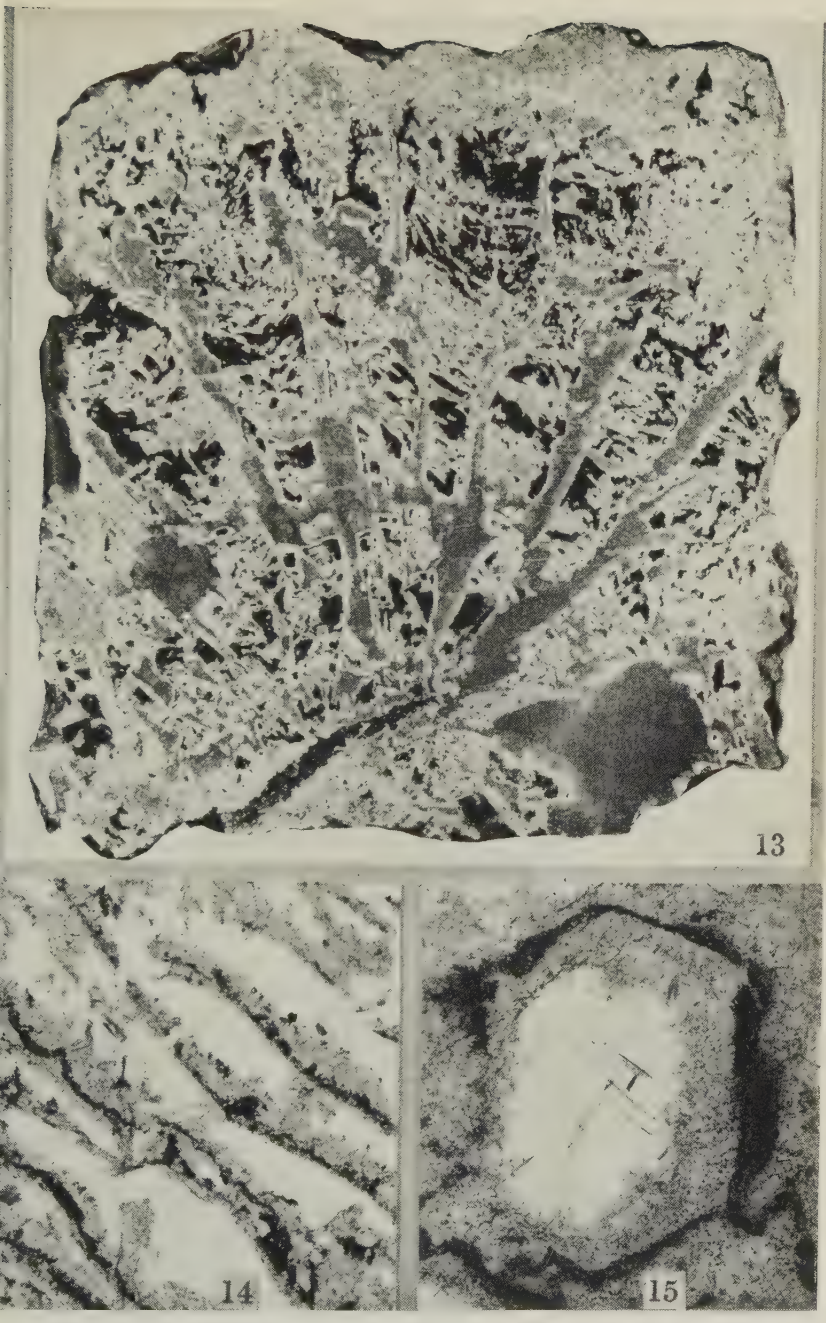
Plate X.

Fig. 16. Anhydrite-dolomite-rock showing replacement of dolomite (dark) by halite which has euhedral tendencies. A thin vein of late halite is also present. Halite-anhydrite zone, 4120–22 feet. ×15.

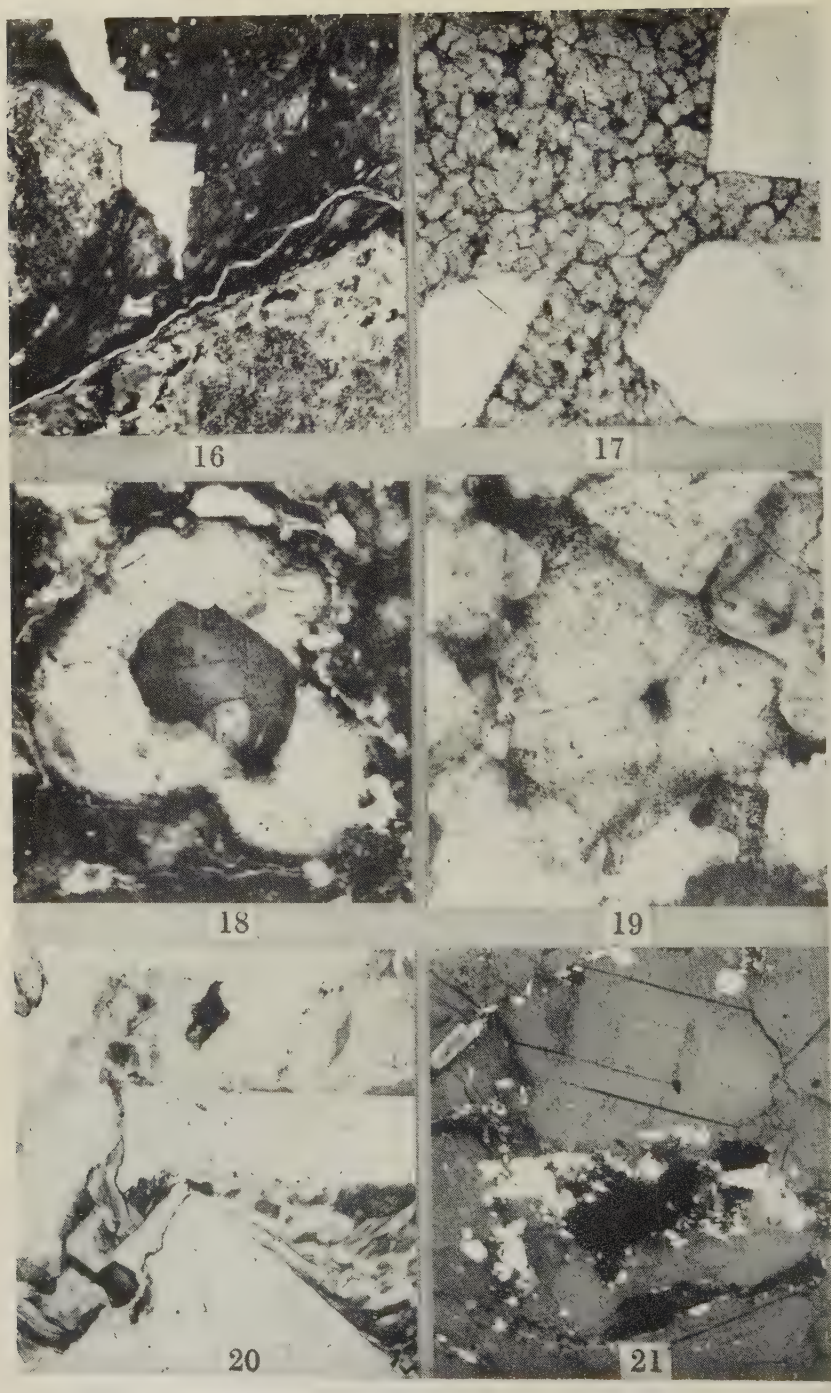
Fig. 17. Halite-anhydrite-rock with fine-grained anhydrite interstitial to large halite grains. Detrital impurities form a honeycomb network in the anhydrite area, probably due to recrystallization of anhydrite from a large number of centres, with ejection of impurities. Halite zone, 3978–80 feet. ×7.

Fig. 18. Halite-sylvine-rock with haematite. A large sylvine grain shows a dark border full of haematite inclusions, and a sharply bounded central area in which haematite inclusions are arranged in two sets of zones inclined to the sylvine cleavages. Dark clay and colourless halite are present. Halite-sylvine zone, 3920–22 feet. ×22.

Fig. 19. Halite-sylvine-rock, showing a large sylvine grain which encloses haematite at its centre and edges. Halite-sylvine zone, 3918–20 feet. ×20.



F. H. STEWART: EVAPORITES, ESKDALE, YORKS.



F. H. Stewart: Evaporites, Eskdale, Yorks.

FIG. 20. Halite-rock with haematite. The dark patches are composed of halite crowded with minute haematite inclusions showing distorted zoning. Upper halite zone, 3916–18 feet. ×7.

FIG. 21. Halite-quartz-rock with clay. The nicol prisms are inclined at an angle of 85° to each other, so that the clay is black and the halite dark. The birefringent grains are mainly quartz, which is associated with the clay. A few grains of magnesite and anhydrite are also present. Upper halite zone, 3910–12 feet. ×15.

Postscript.—Since this paper was written and read to the Society, a paper by Dr. A. Fleck (1950) has been published, and gives details of the results of three recent borings put down by Imperial Chemical Industries in the Yorkshire potash-field. Permission for the publication of the present paper has now, therefore, been given by I.C.I. In Fleck's paper details are given of the thickness and KCl content of two main beds of sylvine and halite: one in the middle evaporite bed and one in the upper evaporites. From the results given it would seem reasonable to conclude that a substantial part of the range from which no cores were recovered in the halite-sylvine zone of the Eskdale no. 2 boring has a high content of sylvine, and not much carnallite. The scarcity of magnesium minerals here is very odd. An upper halite zone is present in the I.C.I. borings, and its very variable thickness adds support to the view that this is a zone of leaching. Fleck mentions the presence of rinneite and a considerable amount of carnallite in the salt clay above the middle evaporites. These would explain the analytical results given on p. 448.

The aegirine-granulites of Glen Lui, Braemar, Aberdeenshire.

(With Plate XI.)

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[Communicated by Dr. S. O. Agrell; read January 26, 1950.]

I. Introduction.

In the explanation of sheet 65 of the Geological Survey of Scotland an outcrop of aegirine-granulites was described from Glen Lui, $1\frac{1}{4}$ miles SSE. of Derry Lodge. Brief descriptions of some of the rocks were given and it was suggested that they might be metamorphosed igneous rocks having distinct affinities with the borolanites of Assynt. Such a correlation, it was felt, would be of importance in assessing the age of metamorphism in the Central Highlands. Harker (1932) made the further suggestion that these rocks might originally have been a tuff which had subsequently become metamorphosed.

Unfortunately, the results of the present investigation do not tend to support these theories and the granulites are regarded as the resultant products of localized metasomatism of the associated metamorphic schists.

Just south of Derry Lodge in Glen Lui, the eastern side of the valley is dominated by a 400-foot hill which rises fairly steeply from the river. It is near the top of this rise that the granulites are located. Across the Glen, about half a mile to the westward is a limb of the Cairngorm granite mass. The intervening ground is unfortunately drift-covered so that the contact between the granite and the Moine gneisses is obscured, but it almost certainly lies to the west of the river. North-east of Derry Lodge this granite has intruded limestones with the production of grossular, idocrase, and wollastonite.

II. Lithological succession.

The aegirine-granulites form a relatively insignificant horizon in the hillside as they extend laterally for some 400 yards and vertically for, at most, 20 feet. Fortunately they are in general well exposed and even

build small crags in places. Several faults tend to complicate the lateral correlation of the beds, but with the aid of one or two well-marked horizons it is possible to form a fairly accurate picture of the succession. An idealized section built up from the better exposed cliffs is given in fig. 1.

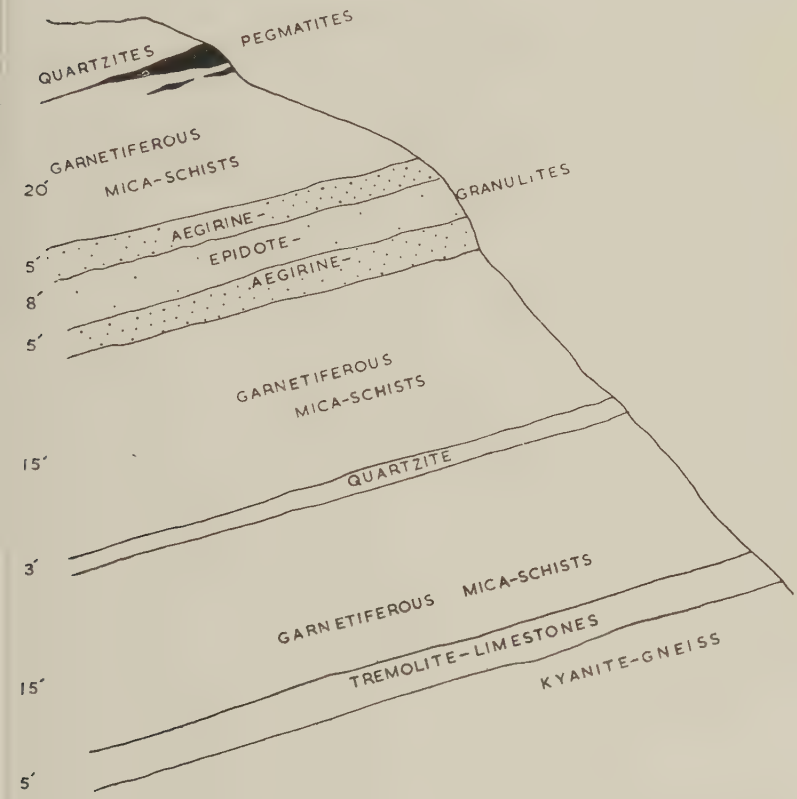

Fig. 1. Idealized section of rock succession in cliffs of Glen Lui, Aberdeenshire.

As the associated metamorphic rocks are of particular interest and have a bearing on the origin of the granulites, they will be described briefly in ascending order.

1. *Kyanite-gneisses.* –The first rocks to appear above the drift-covered lower slopes of the hill are small outcrops of kyanite-gneiss and intercalated limestones and quartzites. The gneisses are coarse, very

acid rocks consisting of quartz lenticles enveloped in felted biotite and containing large porphyroblasts of kyanite or almandine. Their most interesting feature is that they have been subjected to thermal metamorphism which is superimposed on the regional metamorphism responsible for their formation. This is reflected in the alteration of the kyanite porphyroblasts to andalusite, cordierite, and muscovite. A typical example of this polymetamorphic assemblage is shown in pl. XI, fig. 1.

MacKenzie (1949) has described rather similar gneisses from Mull, but he does not mention the presence of cordierite, the change being simply from kyanite to andalusite. MacGregor (1928) found the same alteration to cordierite and andalusite in the aureole of the Lochnagar granite. Tilley (1935) has described the reverse process round the Carn Chuinneag granite where chiastolite suffers replacement by kyanite.

The alteration of kyanite to cordierite suggests that there may have been introduction of Mg and Fe into the gneisses from the granite. Moreover, there are often minute, idiomorphic crystals of tourmaline enclosed in the resultant cordierite and these also suggest the introduction of similar elements. However, tourmaline occurs as a common minor accessory in the main fabric of these gneisses and shows no sign of replacing the other constituents, being present as small rounded crystals. Goldschmidt and Peters (1932) have shown that pelitic sediments may contain up to 0·1 % of B_2O_3, and this would be ample for the formation of the, at most, 1 % of tourmaline present in these rocks. It is therefore almost certain that the small tourmaline prisms in the cordierite were formed by recrystallization of pre-existing tourmaline.

Without introduction of Mg and Fe, the most feasible mode of formation for the cordierite is by reaction between kyanite and its surrounding felt of biotite. Occasionally, as in pl. XI, fig. 1, the kyanite becomes armoured by a mantle of cordierite and is compelled to change directly to andalusite. This strongly suggests that the reaction is local in origin and has not been assisted by elements from the granite. The fact that the biotite in these rocks is usually associated with almandine shows that it is a magnesian type, so that the only constituent necessary is SiO_2 which is present in excess. The following equation requires a change in volume of only 5 % and represents the relative volumes of cordierite and muscovite as closely as can be estimated from thin sections.

$$3H_2O + 7SiO_2 + \underset{\text{(kyanite)}}{8Al_2SiO_5} + \underset{\text{(biotite)}}{2K(Mg,Fe)_3AlSi_3O_{10}(OH)_2} \longrightarrow \underset{\text{(cordierite)}}{3(Mg,Fe)_2Al_4Si_5O_{18}.H_2O} + \underset{\text{(muscovite)}}{2KAl_3Si_3O_{10}(OH)_2}.$$

The kyanite-gneisses are very variable and many carry andalusite which shows no sign of having been derived from kyanite. The more calcareous bands have hornblende and almandine with or without z-oisite and clinozoisite. Large segregations of quartz and kyanite, similar to those described by Read (1933) in the Shetlands, also occur and even in the hand-specimen the kyanite is seen to have pink centres of andalusite. In the Survey Memoir of this area, Barrow (1912) repeatedly states that the intrusion of the 'Newer Granite' masses was not responsible for any marked thermal metamorphism. On the other hand, MacGregor (1928) ascribes similar polymetamorphic phenomena to the intrusion of the Lochnagar mass, and in Glen Lui it seems only reasonable to assume that the same effects were induced by the Cairngorm mass.

Intercalated with the gneisses are narrow horizons of limestone and quartzite, the latter being of particular interest as it often contains abundant octahedra of fluorite particularly in the neighbourhood of faults.

2. *Limestones.*—At about 350 feet above the floor of the valley the kyanite-gneisses give way to some 5 feet of impure limestones. These include bands rich in diopside or tremolite, while others are almost pure marbles.

3. *Micaceous schists and quartzite.*—The next 33 feet are occupied by an almost infinite variety of micaceous schists. The central 3 feet are of pure white quartzite. The micaceous schists are typical medium-grained schists consisting of quartz, blackish-green or brown biotite, muscovite, almandine, and andalusite. Muscovite may form up to 60 % by volume, while biotite may be absent or present with or without muscovite. Almandine is common, usually as small pin-head crystals showing a certain amount of alteration to biotite or chlorite and magnetite. Some of the larger garnets show signs of rotation in the presence of S-shaped lines of iron-ore inclusions. In some bands almandine is present up to 15 % by volume. These schists are of particular importance as they represent types similar to those which appear to have been metasomatized to give the aegirine- and epidote-granulites. Six feet from the top of the garnetiferous schists is an horizon rich in hornblende, and it is particularly useful as a marker because the granulites begin 6 feet above it.

4. *Aegirine-granulites and epidote-granulites.* As shown in fig. 1, the granulites occur as a triple band, the outer two horizons being aegirine-bearing and the centre one more epidotic. In the field the granulites

are white or pink, some retaining a crude foliation while others are more massive felspathic rocks. They are uniformly fine-grained and frequently show narrow transgressive or concordant veins of green aegirine or pale-blue riebeckite-crossite, making their identification in the field quite easy. The granulites are extremely variable and the distinction between aegirine- and epidote-bearing types is by no means clear-cut or always present, indeed in the central area of the outcrop the granulites pinch out to a narrow band of epidote-granulite about 2 feet thick, the place of the other granulites being taken by micaceous schists similar to those described in the last section.

Above the 20 feet or so of granulites there is a return to garnetiferous mica-schists for another 20–25 feet and intercalated with these are flaggy quartzites which become more massive towards the top of the hill.

5. *Soda-pegmatite.*—Intruded into the quartzites in the eastern sector of the exposures is a highly sodic pegmatite which extends laterally for some 30 feet and is up to 3 feet in depth. The exposures are rather poor in this area, however, and the body is probably larger than this.

The pegmatite is composed almost entirely of albite which shows very fine twin-lamellae confined chiefly to the edges of the crystals. In addition there are a few flecks of biotite, chlorite (vermicular pennine), pyrite, and a little tourmaline. The rock is slightly fractured and the fractures are filled first by silica and then by calcite. Since the latter mineral does not appear to have been derived by the alteration of plagioclase and preserves sharp contacts with the albite, a portion of the rock was crushed, treated with dilute hydrochloric acid to remove the calcite, and then analysed (table III, p. 490). As may be seen from this analysis the Na_2O content is high and the ratio Na_2O/K_2O remarkably so. The alkali determination was repeated to minimize any possibility of error.

Judging by the field evidence this pegmatite is not itself a metasomatic product because the quartzites surrounding it have been arched to accommodate the granitic material. The peripheral twinning of the albite crystals also suggests some movement of the body when almost solid. Peripheral twinning is not seen in any of the granulites.

The quartzites surrounding the pegmatite have been albitized for a distance of at least 4 feet above it and in places the contact between igneous and sedimentary material is scarcely distinguishable. The resulting metasomatic product is very like a fine-grained acid igneous rock through which run siliceous replacement veins often carrying pyrite and sphene. Small zircons are also abundant.

The micaceous schists towards the eastern end of the hill are intruded

by numerous small veins and lenticles of albite and in the field these appear very similar to the main, larger body of pegmatite. One of the small bodies in particular merits fuller description. It is injected into an acid muscovite-biotite-schist about 10 feet below the general level of the granulites. As a result of the injection of the felspathic material an area about 1 foot by 6 inches in size has become highly felspathized with

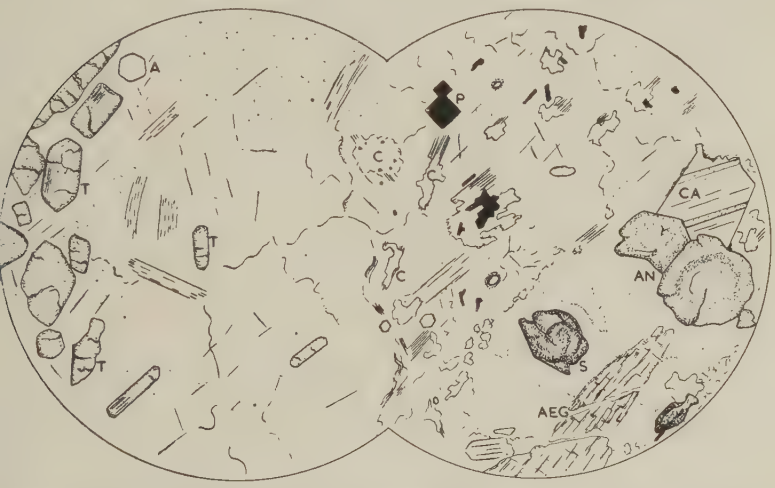

Fig. 2. Aegirine-augite (Aeg) and andradite (An) produced in a micaceous schist by the intrusion of a tourmaline (T)-bearing pegmatite. The groundmass is mainly albite with apatite (A), chlorite (C), pyrite (P), calcite (Ca), and sphene (S). Muscovite and zircon are also present.

the formation of albite, andradite, aegirine-augite, sphene, and abundant tourmaline (fig. 2). The presence of dense tourmaline aggregates shows that this vein is not a simple metamorphic segregation, so that we have here an example of aegirine-augite developed in a micaceous schist by the intrusion of pegmatitic material. Other much narrower veinlets of plagioclase and tourmaline intrude the schists for some distance from the pegmatite, up to 50 yards or so, but their effect on the schists is usually confined to felspathization for an inch or two on either side.

The intrusion of the micaceous schists by these veinlets of albite has frequently produced a texture which can only be described as micro-migmatitic, and throughout the granulites themselves the same texture persists though often on a microscopic scale. This is particularly true of the aegirine-granulites, the epidote-bearing varieties being usually more homogeneous rocks.

Petrography of the granulites.

Petrographically the granulites display an almost bewildering variety of types ranging from slightly felspathized mica-schists to aegirine- and epidote-bearing highly felspathic granulites. Every gradation between these three types exists and may even be present within the compass of one slice.

1. *Felspathized mica-schists.*—These are schists in which orthoclase and albite have been developed at the expense of the original quartz and mica. Folia of biotite and muscovite, particularly the latter, may be seen to fade into orthoclase, so that the original extent of the mica can only be inferred from the lines of black flecks extending into the felspar. The grain-size of these rocks and the granulites averages about 0·25–0·5 mm., so that they are fine-grained except in the pegmatitic patches where the felspars may be up to 1 or 2 cm. in length.

As the amount of felspar developed increases, minerals such as andradite, aegirine-augite, epidote, and sphene begin to appear and the schistosity is lost, giving rise to the following rock types.

2. *Aegirine- and epidote-granulites.*—The main fabric of the aegirine- and epidote-granulites consists once again of a granulitic fine-grained groundmass of orthoclase and albite, while quartz has usually disappeared completely. The texture formed by the felspars may be judged by the groundmass of pl. XI, fig. 2. Staining of a typical epidote-granulite with sodium cobaltinitrite after the method of Gabriel and Cox (1929) showed that the potash and sodic felspars occurred in definite layers. Moreover, when the bands rich in potash were traced across the schistosity into less felspathic areas of the specimen, the corresponding bands were found to be rich in muscovite. This restriction of the orthoclase to definite layers suggests very strongly that it was formed by metasomatism of muscovite, since it is difficult to imagine the original igneous rock or tuff being intimately stratified into bands alternately rich in soda and potash, two or three to the centimetre.

The albite (An 7–11) between the bands of orthoclase is pink in the hand-specimen and intensely clouded in thin section. It is seldom twinned except in some of the aegirine-granulites, where it occurs as clear well-twinned crystals forming minute veins and pegmatite-like patches.

The potash-felspar was described in the Memoir of the Geological Survey (1912) as 'an alkali-felspar containing both potash and soda, and therefore probably belonging to the anorthoclase group'. This statement was presumably based on the low optic axial angle of these felspars, as the slide 8217 described in the Memoir contains potash-felspar with

a 2V of 54° measured on the universal stage. Similar values for the aegirine- and epidote-granulites analysed and given in table III are 51° and 49° respectively. In the more migmatitic aggregates the 2V may rise to 65°, and in these cases the felspar is distinctly perthitic, whereas the small crystals of the groundmass with lower 2V values extinguish almost perfectly uniformly.

A conspicuous feature of the aegirine-granulites of the cliffs nearest Derry Lodge is a line of large porphyroblasts of potash-felspar. These may reach a length of 15 cm., are white in colour, and have been replaced peripherally and along cracks by pink albite. The 2V of these porphyroblasts is 58°, so that they are comparable with the smaller felspars of the groundmass. Moreover, the fact that they are not microcline suggests that they are products of the thermal metamorphism and metasomatism rather than relict crystals from the Moine schists. Since E. Spencer (1937) has shown that potash-felspars of the orthoclase-microperthite series may have low optic axial angles, one of these large porphyroblasts was analysed to determine the ratio of K_2O to Na_2O. This was found to be 14·9 : 1·4 %, so that allowing for a probable 1·5 2·0 % of anorthite, the content of soda-lime-felspar would not exceed 12 %. This is in fairly close agreement with the curves drawn up by E. Spencer (1937), and since Alling (1921) restricts the term anorthoclase to types containing over 30 % of albite, the term orthoclase will be used here. A careful comparison of the modes and chemical analyses of the granulites does not indicate any large amount of soda in the orthoclase as there is usually insufficient potash-felspar in the mode to satisfy the potash of the analysis. This discrepancy is due in some cases to the presence of muscovite and in others more probably to the difficulty in distinguishing small crystals of orthoclase from untwinned albite.

A feature common to both the aegirine- and epidote-granulites is that they frequently carry two distinct garnets. It is often difficult, in thin section, to determine the exact types of garnet involved, and to prove the existence of two types a typical felspathic granulite was crushed and the heavy residue above 3·6 was further fractionated in Clerici solution of sp. gr. 3·85. This process yielded 0·4 gm. of a pink almandine and 1·6 gm. of a yellowish andradite. In view of the very small amount of almandine recovered, further purification of this mineral was not attempted. Both garnets were analysed (table I), but the analysis of the almandine is not too satisfactory, owing to the small amount available and its probable impurity. The significant fact, however, in the present discussion is that there are clearly two distinct garnets present in the same rock.

TABLE I. Chemical analyses of two garnets from the same rock (granulite) from Glen Lui, Aberdeenshire.

1. Almandine (Analyst, J. H. Scoon). 2. Andradite (Analyst, G. R. McLachlan).

	1.			2.		
SiO_2	39·69	3·10	3·10	36·2	2·99	2·99
Al_2O_3	20·59	1·89	1·89	4·95	0·49	1·95
Fe_2O_3	2·77	0·17		23·25	1·46	
FeO	24·06	1·57		1·33	0·09	
MnO	2·64	0·17	2·80	0·43	0·03	
MgO	1·07	0·12		—	—	2·99
CaO	8·37	0·70		30·86	2·74	
TiO_2	1·25	0·07		1·88	0·13	
H_2O— ...	0·21			0·33		
	100·65			99·23		
Sp. gr. ...	4·13			3·76		
n	1·807			1·87 approx.		

In some cases the difference between the two garnets is quite obvious, as yellowish andradite may be seen moulded on colourless cores of almandine (pl. XI, fig. 2). The andradite is then fresher than the almandine which is cracked and frequently altered to biotite and chlorite. Occasionally the andradite may form large ragged porphyroblasts, but it is then, in part at least, anisotropic, yellow, and not altered to biotite and chlorite.

Andradite also occurs in the aegirine-granulites as minute idiomorphic crystals, which, with the pyroxene, may form long winding trails through the rocks. These small crystals are pale yellow, generally isotropic, but often marginally anisotropic. Larger crystals up to 2 mm. are characteristic of the felspathic veins near the pegmatite, and these are uniformly anisotropic and often show beautiful sector twinning. Some are heavily zoned, presumably with varying tenor of the grossular molecule, and this forms a distinguishing feature when almandine is also present.

The coexistence of two distinct garnets, one characteristic of acid metamorphic schists and the other more typical of contact metamorphic and replacement deposits, is probably the most significant feature of the granulites, as it is difficult to imagine their being produced by metamorphism of an igneous rock or tuff. On the other hand, if the granulites are regarded as the result of metasomatism of an almandiferous schist, the almandine may be regarded as a relict mineral. The virtual identity of the almandine of the granulites with that of the underlying schists is shown by the similarity of their refractive indices. That of the former is 1·807 and of the latter 1·811.

The greater persistence of the almandine in the micaceous bands of

the granulites is in support of its inferred relict nature, while the frequent association of andradite with aegirine-augite in the felspathic veins shows that the calcic garnet is the stable mineral under the metasomatic conditions.

Epidote is a common constituent of the granulites, occurring as small sub-idiomorphic crystals, colourless or pale yellow, with 2V negative about 85°. It reaches its greatest development in the felspathic veins near the pegmatite and the crystals may reach a length of 6 mm. Epidote is not often associated with aegirine in the granulites, the significance of which will be dealt with later. In general the two minerals characterize different bands, but a few grains of diopside may be present in the epidote-granulites.

The aegirine-augite of the granulites occurs in three distinct forms. Firstly, as interstitial allotriomorphic masses which often extend laterally and form pyroxene-rich bands parallel to the general schistosity of the outcrop. Secondly, the mineral appears as sub-idiomorphic skeletal crystals up to 3 mm. in length which often grow round aggregates of chlorite, felspar, or foxy-red biotite. Invariably when the crystals are of this type they are associated with very clear twinned albite and quartz which may project as perfect crystals into cavities later filled by calcite. A typical example of these skeletal forms is represented in pl. XI, fig. 3.

The third mode of occurrence of the pyroxene is as small, often minute, idiomorphic prisms which are once again associated with clear albite. The dark-green, often radiating acicular crystals may be scattered at random throughout the albite, but more often they tend to form aggregates or dense bands parallel to the general schistosity of the granulites. It is also as needles of this sort that the aegirine-augite is present with andradite in narrow trains ramifying through the rocks.

Alteration products of the pyroxene include a pale-blue sodic amphibole (riebeckite-crossite), while inclusions of magnetite are characteristic of the denser masses. The sodic amphibole may either form peripheral fringes on aegirine-augite, or it may be found independently as fine tufts which combine to form narrow bands as with the pyroxene. It also occurs as hair-like tufts in quartz. When some of the granulites are cleaved, the schistose surfaces may appear quite blue from the abundance of riebeckite-crossite, so that they may be easily recognized in the field. The amphibole, however, is not as common a mineral as the pyroxene, but like the latter it attains its maximum development in association with large veinlets of albite, the strings of riebeckite-crossite often outlining the felspathic body as if forced aside by the growing albite.

Both the pyroxene and the amphibole were separated and analysed; the analyses together with the optical data are given in table II. The intense colour of the amphibole made it impossible to assess the 2V accurately.

TABLE II. Chemical analyses and optical data of pyroxene and amphibole from granulite, Glen Lui, Aberdeenshire.

1. Aegirine-augite (Anal. G .R. M.). 2. Riebeckite-crossite (Anal. G. R. M.).

	1.				2.		
	Wt. %.	No. of metal atoms to 6 oxygens.			Wt. %.	No. of metal atoms to 24(O,Oh,F)	
SiO_2 ...	51·68	1·961		2·000	51·02	7·499	7·834
Al_2O_3 ...	1·28	0·059	0·039		1·93	0·335	
			0·020				
TiO_2 ...	1·59	0·045			1·41	0·158	
Fe_2O_3 ...	19·10	0·546		1·033	13·05	1·446	
FeO ...	5·17	0·164			10·96	1·340	4·848
MnO ...	0·45	0·014			0·14	0·009	
MgO ...	4·29	0·244			8·62	1·895	
CaO ...	9·56	0·389		0·927	4·85	0·776	
Na_2O ...	7·25	0·538			4·95	1·429	2·328
K_2O ...	—	—	—	—	0·74	0·123	—
H_2O+ ...	0·05	—	—	—	2·43	2·346	
H_2O- ...	0·05	—	—	—	0·25	—	2·434
F	—		—		0·2	0·088	
	100·47				100·55		

Sp. gr. (20° C.)	3·561	3·192
α	1·741 deep green	1·668 pale blue
β	1·767 yellowish-green	— yellowish
γ	1·789 brownish-green	1·680 bluish-mauve
$2V_\alpha$	85°	50° ± axial plane
$\alpha : c$	12°	14° ⊥ (010)

The aegirine-augite is homogeneous and never shows any sign of zoning beyond a slight irregular patchiness of colour in the larger crystals of the allotriomorphic aggregates.

Sphene is a very common accessory in all the Glen Lui rocks and shows considerable variation in colour and mode of occurrence. In the granulites it varies from small colourless blebs to typical greyish lozenge-shaped crystals and aggregates. When it occurs in the clearer quartz-albite pegmatitic veinlets in association with aegirine-augite, it often shows reddish patches around what appears to be an earlier core on which more idiomorphic crystals have moulded themselves (fig. 3).

Many of the granulites are very rich in small rounded crystals of zircon. These are often concentrated along definite horizons and even in one slice they may be traced through the section in lines parallel to the general schistosity. Their abundance, rounding, and stratification suggested that they might be relict sedimentary crystals. To establish this on a more quantitative basis, the length/breadth ratios of 200 crystals from a felspathic granulite were plotted against similar ratios for 200 crystals from an underlying garnet-muscovite schist. These values are given in fig. 4 according to the method of Poldervaart and von Backström (1950) and Poldervaart (1950). The close similarity of the curves with maxima at a point between 1·65 and 1·85 suggests that the zircon in both rocks has been subjected to a comparable degree of rounding. Moreover, the curves are fairly similar to those obtained by Poldervaart (1949, 1950) for schists and gneisses in the Kakamas area.

Fig. 3. Aegirine-augite in micropegmatite with sphene showing two stages of growth.

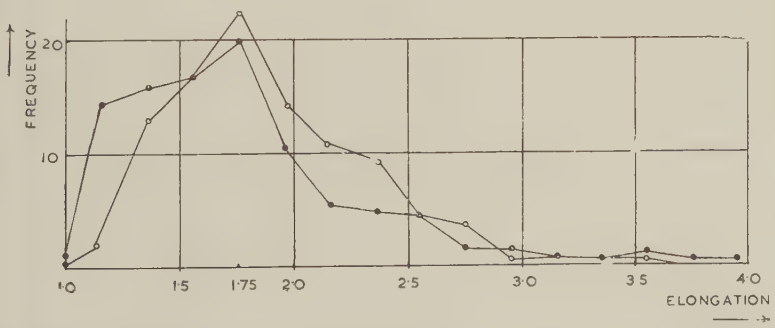

Fig. 4. Percentage frequency of zircon length/breadth ratios from a felspathic granulite (solid dots) and an underlying garnetiferous muscovite schist (open dots).

Reynolds (1936) has shown that zircons in xenoliths may retain their sedimentary characteristics during metasomatism, so that in Glen Lui the rounded crystals may be regarded as relict minerals from the original schists. If, on the other hand, the granulites were formed by metamor-

phism of an igneous rock, the zircons should show igneous features and the same would be true if they had crystallized from metasomatizing fluids. A tuff might be expected to show some rounding of its zircons, but in view of the restricted field distribution of the Glen Lui rocks, it is unlikely that the tuffaceous material could have been transported far, and this would have prevented the zircons from suffering any considerable attrition.

Some of the muscovite in the granulites, particularly that in the neighbourhood of the tourmaline veins, is clearly of metasomatic origin. Much of it, however, preserves a distinct schistosity identical with that of the associated schists. Aegirine-augite is frequently found associated with muscovite and quartz, and this assemblage would be difficult to imagine in a purely metamorphic rock, since aegirine is characteristic of peralkalic rocks and muscovite of peraluminous acidic types. It is extremely unlikely that an igneous rock or tuff would give rise to such a combination even under metamorphism. If, however, it is accepted that the granulites are metasomatic in origin, then the persistence of muscovite as a relict mineral is merely dependant on the ability of the sodic metasomatizing agents to convert it to felspar. Failing such a conversion, or subsequent to it, the soda would enter the pyroxene.

The biotites of the granulites are usually dark-brown or green varieties variously bleached and altered to yellowish and green chlorites, some showing anomalous and others normal interference colours. That these chlorites are derived from biotite is shown by their inclusions of minute octahedra of magnetite, which may congregate in masses or form long winding strings with chlorite. Apart from these types, secondary after biotite, there are large interstitial masses of pale green pennine often beautifully vermiculate or gathered into rosettes and showing a distinct tendency to be associated with andradite. Identical material is present in the pegmatite and associated veinlets and in most cases is clearly of late formation, as the aggregates are bounded by sharp contacts against the surrounding minerals. Apatite is another common accessory of the granulites and may be present in amounts up to 1 % by volume.

3. *Silicified granulites.* Many of the aegirine-granulites show unmistakable signs of silicification as the last stage in their formation. This is most pronounced in the bands of clear albite and aegirine-augite and of some cavities in the granulites which have become filled with quartz or calcite into which albite or pyroxene crystals project. As might be expected, andradite is unstable under these conditions and the idiomorphic crystals are replaced by iron-ore as is shown in pl. XI, fig. 4.

Aegirine-augite may also disappear, so that its former presence can only be inferred from the 'ghost' outlines of black flecks which remain in the quartz. Two of the 'ghosts' may be seen in pl. XI, fig. 4 near the upper left-hand corner.

The silicification of the granulites is most pronounced in the horizon at the top of the upper aegirine-granulite band and closely resembles that produced in the quartzites in contact with the soda-pegmatite. In both cases the silicification is clearly the last stage of the metasomatism, replacing all the previously formed minerals.

Before considering the chemical implications of the metasomatism under discussion it might be as well to summarize the purely petrographic evidence in favour of a metasomatic origin for the granulites.

(1) The great mineralogical and textural variety of the granulites. Had they been formed by metamorphism of an igneous rock or tuff, greater uniformity might have been expected.

(2) The presence of the pegmatite shows that a source of soda was available.

(3) The fact that tourmaline-bearing veins have produced aegirine and andradite in the underlying schists shows that the mineral assemblage of the granulites could be produced in those schists.

(4) The presence in felspathic granulites of two garnets, one similar to those in the associated schists and the other typical of contact metamorphic rocks.

(5) The occurrence of the orthoclase in definite bands which appear to correspond to the muscovite folia of the schists.

(6) The coexistence of muscovite and aegirine, the former characteristic of peraluminous and the latter of peralkalic rocks.

(7) The widespread dissemination of hydatogenic minerals, characteristic of hydrothermal replacement, throughout the granulites and associated schists and felspathic veins. These minerals include tourmaline, fluorite, pyrite, andradite, and aegirine-augite.

(8) The sedimentary nature of the zircons, and the comparable degree of rounding shown by the zircon of the granulites and the underlying schists.

Chemistry of the metasomatism.

Having considered the petrographic evidence in favour of metasomatism, it is now necessary to decide the type of metasomatism involved. In table III, six analyses are given. The first is that of the soda-pegmatite which is composed almost entirely of albite (An 3 %), some 15 % of

quartz, largely in the form of narrow veins filling cracks between the felspar crystals, and about 4 % of sericite, tourmaline, and pyrite. This is reflected in the analysis since all the values except SiO_2, Al_2O_3, and Na_2O are less than 0·7 %. The extremely sodic nature of this pegmatite suggests that it was intruded very late in the differentiation of the Cairngorm mass.

TABLE III. Chemical analyses and mineral composition of rocks from Glen Lui, Aberdeenshire.

	1.	2.	3.	4.	5.	6.
SiO_2	69·94	62·72	61·79	56·84	67·77	60·70
Al_2O_3	19·02	17·33	16·90	18·78	16·42	19·79
Fe_2O_3	0·62	3·23	3·10	3·98	1·37	3·63
FeO	0·38	1·30	1·07	2·76	3·68	3·63
MgO	0·01	0·90	0·90	1·22	1·50	0·98
CaO	0·46	1·75	2·44	3·84	1·23	0·68
Na_2O	8·99	5·87	5·26	4·75	2·28	0·42
K_2O	0·52	4·48	6·96	6·26	3·58	6·44
H_2O+	0·34	0·60	0·29	0·72	0·74	1·70
H_2O-	0·02	0·16	0·14	0·17	0·04	0·18
TiO_2	0·03	0·97	0·90	0·39	1·04	1·32
P_2O_5	—	0·15	0·34	0·47	0·15	0·15
MnO	0·12	0·04	0·19	0·04	0·33	0·18
	100·45	99·50	100·28	100·22	100·13	99·80
$\frac{Na_2O+K_2O}{Al_2O_3}$	0·81	0·84	0·96	0·78	0·47	0·38
Sp. gr.	2·58	2·65	—	2·72	—	—
Quartz	15·0	—	—	—	—	—
Orthoclase	—	26·0	35·0	35·0	—	—
Albite	81·0	53·0	45·0	47·0	—	—
Muscovite	3·0	8·5	—	—	—	—
Aegirine-augite ...	—	7·0	9·0	—	—	—
Riebeckite-crossite ...	—	3·0	—	—	—	—
Almandine	—	—	6·0	—	—	—
Andradite	—	0·5	3·0	—	—	—
Epidote	—	—	—	9·0	—	—
Sphene, ores, apatite, &c.	1·0	2·0	—	6·5	—	—
Chlorite	—	—	2·0	2·5	—	—
	100·0	100·0	100·0	100·0		

1. Soda-pegmatite from the upper quartzites of Glen Lui. (Analyst, G. R. M.)

2. Aegirine-granulite from the lower granulite contact, Glen Lui. (Analyst, G. R. M.)

3. Aegirine-granulite, no. 8217, Mem. Geol. Surv. Braemar, 1912. (Analyst, W. Pollard.)

4. Epidote-bearing felspathic granulite. Glen Lui. (Analyst, G. R. M.)

5. Garnetiferous muscovite-biotite-schist, Glencalvie Lodge, Mem. Geol. Surv. Ben Wyvis, Ross-shire, 1912, p. 112. (Analyst, W. Pollard.)

6. Mica-schist, Banffshire, Mem. Geol. Surv. Banffshire, 1923. (Analyst, E. G. Radley.)

The second analysis is of a felspathic granulite with some persistent schistose lamellae of muscovite and biotite, while aegirine-augite, riebeckite-crossite, and andradite are also present. No. 4 is of a homogeneous very felspathic epidote-granulite, once again with some micaceous bands.

Analysis 3 is from the Memoir of the Geological Survey, slide 8217. It is a typical aegirine-granulite with both almandine and andradite. Accompanying the analyses are the percentages by weight of the component minerals. In some cases these may not appear to correspond too well with the analyses. This is in part due to the banded nature of the granulites, so that the slide on which the mode is based may not cover parts of the specimen which are crushed for analysis.

Numbers 5 and 6 were chosen from analyses of Scottish schists to show the approximate chemical composition of the Glen Lui garnetiferous mica-schists. Since they vary from biotite-muscovite-schists to others containing 20–40 % of muscovite, no. 5 should represent the former accurately enough, while no. 6 is higher in K_2O and Al_2O_3, consistent with a higher percentage of muscovite. Since some of the Glen Lui schists contain small amounts of andalusite, Al_2O_3 probably reaches 20 % and may be down to 14 % in the less micaceous bands. The Al_2O_3 of the granulites also lies roughly between those two values, so that it seems unnecessary to postulate introduction of alumina. The values for K_2O, varying from 3·7 to nearly 7·0 %, would also be covered by the muscovite content of the original schists and, in addition, the almost complete absence of K_2O in the pegmatite suggests that the metasomatizing fluids did not contain any significant amounts of potash.

The status of calcium is more difficult to assess since its low value in the pegmatite makes its introduction unlikely, but the granulites contain up to 3·84 % of CaO. That the associated schists contain a certain amount of calcium is shown by the fact that in a narrow vein of felspar and tourmaline intruded into them, the albite has become basified to oligoclase (An 28). The bands of tremolite also point to the presence of calcium in many of the schists. It is therefore unnecessary to assume introduction of calcium. The calcite which is present both in the pegmatite and the granulites may be ascribed to the circulation of groundwater.

The only major constituent which appears to have been introduced, or at any rate fixed, in the granulites is sodium, since the Na_2O content of the original schists is unlikely to have exceeded 1·5 %, whereas in the granulites it rises as high as 5·87 and does not fall below 4·75 %.

It would seem as if the granulites were formed by some process such as the following. During the intrusion of the Cairngorm granite the schists on the eastern side of Glen Lui were subjected to thermal metamorphism sufficiently intense to induce the conversion of kyanite to andalusite and cordierite. In the later stages of the intrusion a highly sodic pegmatite was injected into the quartzites while associated sodic solutions permeated the underlying schists parallel to their schistosity. By hydrolysis of biotite and muscovite, Na became fixed with the formation of albite, the K of the micas similarly forming orthoclase. As the available alumina became exhausted the Na was compelled to crystallize as aegirine-augite and riebeckite-crossite with Mg and Fe derived from the hydrolysis of biotite. Moreover, as the temperature fell and the metasomatizing power of the solutions was reduced, muscovite was no longer converted to felspar, and soda-pyroxene was formed in spite of the presence of the mica.

Epidote might be expected to form when alumina was available after the formation of the felspars and provided calcium was also present. A schist rich in muscovite would provide these conditions and the banding of the orthoclase in the epidote-granulites and the persistence of some muscovite-rich layers show that the epidote-granulites were formed from schists of this sort.

All the Glen Lui analyses show a high ratio of Fe_2O_3 to FeO, and this is particularly noticeable in the riebeckite-crossite and the aegirine-augite. The formation of andradite may be attributed to this change in the state of oxidation of the schists, while paucity of alumina kept the grossular content of the garnet low. The fact that calcium was available shows that under the particular conditions prevailing, an almost pure albite was the stable plagioclase and little calcium entered into its formation.

There is an important difference in the mode of replacement of the two types of granulite. In the epidote-granulites the replacement is on a volume-for-volume basis, the resulting product being a dense homogeneous rock, whereas in the aegirine-granulites the replacement has been accompanied by a decrease in the specific volumes of the minerals with consequent cavernous replacement. As a result the aegirine-granulites are porous and frequently contain cavities filled later by silica or calcite. Furthermore, the epidote-granulites have not been silicified, so that the aegirine-bearing rocks may be said to have a longer metasomatic history than the others.

According to Turner (1948) soda-pyroxenes may be expected in a

metamorphic rock when the ratio $Na_2O + K_2O/Al_2O_3$ approximates unity. The values for this ratio in the Glen Lui rocks are given below the analyses on page 490, but the results are somewhat anomalous. Aegirine-augite here begins to appear when the ratio is only 0·84, but this is due to the fact that aegirine-augite and muscovite are found together in this particular rock. As a result, had complete reaction occurred between the sodic solutions and the muscovite, aegirine-augite would not have appeared. In this way its formation at a ratio of 0·84 may be explained, whereas in stable rocks it would probably not appear until a ratio of 0·90–0·95 had been reached.

Since the granulites follow a definite horizon in the succession and do not occur as random patches scattered throughout the schists, it seems necessary to assume that a pegmatite is either concealed deeper in the hillside at the same horizon, or else that one has been removed by erosion on the side of the hill towards the river. The band of silicification at the top of the granulites suggests this very strongly, as a similar effect may be observed in the neighbourhood of the pegmatite intrusive into the quartzites.

Examples of soda-metasomatism are common enough in the literature, but none can be said to be identical with the situation at Glen Lui. Goldschmidt (1914) attributes the formation of riebeckite and aegirine-diopside in hornfels to the introduction of soda from the larvikites and nepheline-syenites of Langesundfjord. The same author (1921) in the Stavanger area describes the formation of albite porphyroblasts in phyllite by the introduction of soda, but here, although the Al_2O_3 percentage is of the same order as in Glen Lui (15–20 %), the soda content does not rise above 3·1 %, so that aegirine-augite is not produced. Taliaferro (1943) shows that the glaucophane-schists of California have been produced by introduction of soda, iron, magnesia, lime, and alumina into sandstones, shales, and cherts. Suzuki (1934) describes quartzose rocks from Hokkaido in which soda-metasomatism has produced schistose bands of aegirine-augite and glaucophane, but here the rocks are so siliceous that Al_2O_3 only rises to 2·73 %, so that the amount of soda introduced is very small.

The work of King (1943) in the Cnoc nan Cuilean area forms the closest parallel to Glen Lui. Here, according to King, siliceous granulites have been metasomatized to hybrid syenites containing anorthoclase, albite, aegirine-augite, and sodic hornblendes. The effects at this locality, however, are on a much larger scale and there is more intimate mixing of the igneous and metamorphic material than in Glen Lui.

The material described in the foregoing pages was largely collected by Dr. S. O. Agrell and this was supplemented by the present writer in two short field trips. The writer wishes, in particular, to record his thanks to Dr. Agrell for allowing him to work on this problem and for considerable advice and assistance during its elucidation.

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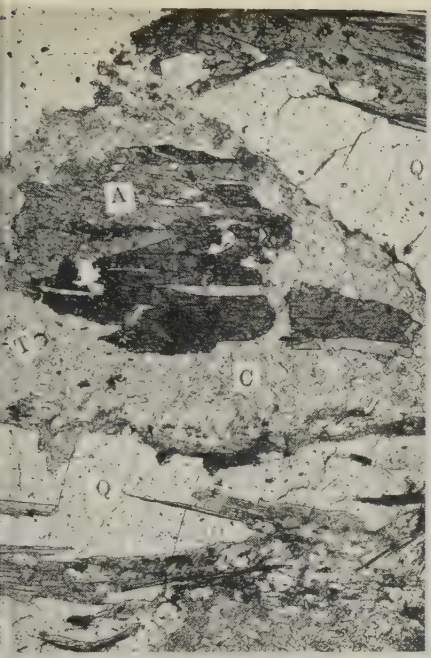

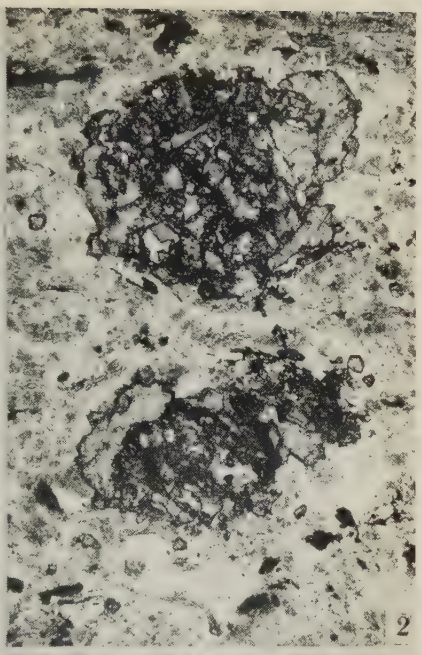

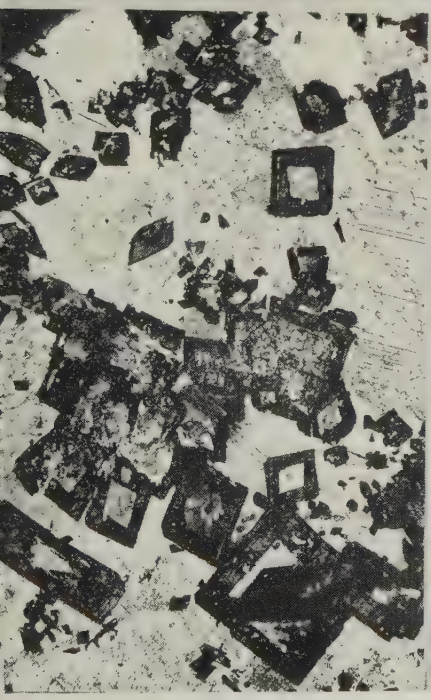

G. R. McLachlan: Aegirine-granulites, Glen Lui, Aberdeenshire

Explanation of Plate XI.

Photomicrographs of aegirine-granulites from Glen Lui, Braemar, Aberdeenshire.

Fig. 1. Kyanite (dark) altered to andalusite (A) and surrounded by cordierite (C) and muscovite. Quartz (Q) and tourmaline (T). ×23.

Fig. 2. Yellowish andradite moulded on almandine. Ordinary light. ×32.

Fig. 3. Aegirine-augite forming skeletal crystals round foxy-red mica. Calcite at right. Ordinary light. ×27.

Fig. 4. Andradite and aegirine-augite in silicified granulite. Andradite pseudomorphed by haematite and limonite. Aegirine-augite appearing as 'ghost' crystals. Ordinary light. ×27.

Priderite, a new mineral from the leucite-lamproites of the west Kimberley area, Western Australia.

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[Communicated by Prof. R. T. Prider; read November 2, 1950.]

Introduction.

During an analysis by X-ray diffraction techniques of some fine-grained leucite-lamproites from the west Kimberley area of Western Australia the author became interested in the rutile in these rocks. Although rutile had been repeatedly observed under the microscope (1, 2), its presence could not be confirmed using X-ray diffraction methods. At the author's request, Professor R. T. Prider supplied a small amount of the mineral which had been identified as rutile. A study of this mineral showed it to be a new mineral similar to rutile in its optical properties.

Priderite is the name suggested for this mineral. Rex Tregilgas Prider, Professor of Geology at the University of Western Australia, has contributed much to the knowledge of Western Australian rocks and minerals, and in particular he has studied in detail the suite of rocks from which this mineral was separated.

General description.

A mineral similar to rutile in its properties is a constant accessory in all the leucite-bearing rocks of the west Kimberley district of Western Australia (1, p. 50). This mineral has, in earlier papers, been referred to as rutile. In the finer-grained lamproites (fitzroyite, cedricite, and mamilite) it occurs as minute reddish rods of the order of 0·05 mm. long, but in the coarser wolgidites (1, pp. 53 and 58) it occurs as stout prisms measuring up to $1 \times 0{\cdot}5$ mm. These 'rutile' prisms exhibit a peculiar lamellation parallel to the base, and in the rocks from Wolgidee Hills this is seen to be due to the crystals being built up of 'rhombic-shaped plates' (1, p. 58). This so-called 'rutile' was separated from the Wolgidee Hills type specimen of wolgidite (which also yielded the two other new minerals, wadeite and magnophorite, described by Prider (2)) for chemical and X-ray work. An examination of this material in oils

reveals that when crushed it breaks into extremely thin nearly rectangular cleavage flakes—the 'rhombic shape' of the flakes mentioned by Wade and Prider (1, p. 58) being due to the fact that the sections in which they were seen were cut somewhat obliquely to the base. This mineral has been re-examined by R. T. Prider and the characteristics are as follows:

Habit: rectangular prisms with a lamellation parallel to the base.

Cleavage: basal highly perfect, and fair parallel to the prism, so that it breaks into extremely thin rectangular plates.

Colour: black with adamantine lustre (similar to nigrine); streak, grey.

Optical characters: very thin basal plates are yellowish-brown and show the emergence of the optic axis normal to the cleavage; uniaxial with positive sign; very high refractive index, ω greater than 2·10. Does not fluoresce in ultra-violet light of wave-length 2537 Å. With these characters the mineral therefore resembles rutile very closely.

Pleochroism: in thin section it is deep reddish-brown with ω reddish-brown, ϵ deep reddish-brown to black.

The crystal fragments used in the X-ray study were 0·1–0·3 mm. in their greatest dimension. They were weakly attracted by an electromagnet indicating that they are paramagnetic. The density determined on 0·2 gram of priderite using a micropyknometer similar to that described by Winchell (3) gave a value of 3·86±0·08 g./cm.^{3} Priderite was unchanged, as judged by its X-ray diffractions pattern, after heating to 1400° C.

The result of a chemical analysis made on a sample of 0·15 gram of priderite is shown in table I.

TABLE I. Chemical analysis of priderite.

			Unit-cell contents.			
SiO_2	...	0·0				
TiO_2	...	70·6	Ti^{4+}	...	6·48	7·95
Fe_2O_3	...	12·4	Fe^{3+}	...	1·14	
Al_2O_3	...	2·3	Al^{3+}	...	0·33	
BaO	...	6·7	Ba^{2+}	...	0·32	1·33
K_2O	...	5·6	K^+	...	0·87	
Na_2O	...	0·6	Na^+	...	0·14	
CaO	...	trace	O	...	16	
MgO	...	0·0				
		98·2				

The analysis was performed by the Assay Department of the School of Mines, South Australia, and J. D. Hayton of the Division of Industrial Chemistry, Commonwealth Scientific and Industrial Research Organization.

TABLE II. Powder diffraction patterns of priderite, Ba-priderite, and cryptomelane.

Priderite			Ba-priderite			Cryptomelane[1]	
d/*n* (obs.) *kX*.	*I*.	*hkl*.	*d*/*n* (calc.)	*d*/*n* (obs.) *kX*.	*I*.	*d*/*n* (obs.) Å.	*I*.
7·13	m	110	7·135	7·12	vw.d	6·92	w
5·04	m	200	5·045	5·06	w	4·91	w
3·57	vw	220	3·567	3·55	w	3·47	vw
3·19	s	310	3·190	3·19	s	3·11	m
2·845	vvw	101	2·841	2·817	vwd	—	—
2·527	vw	400	2·523	—	—	2·46	vw
2·470	m	211	2·475	2·470	m–s	2·40	s
2·379	vvw	330	2·378	—	—	2·325	(w)
2·258	vw	420	2·256	2·250	vw	2·21	w
2·222	w	301	2·222	2·223	m	2·16	w
2·034	vw	321	2·034	2·032	vwd	—	—
1·981	vw	510	1·979	1·977	vw	1·935	(m)
1·887	m	411	1·886	1·884	w–m	1·835	m
1·787	vvw	440	1·783	—	—	1·74	(w)
1·737	vvw	530	1·730	—	—	1·69	(w)
1·684	w	600	1·682	1·683	m	1·64	w
1·585	w–m	620 521	1·595 1·579	1·583	w–m	1·55 1·54	(w) m
1·480	w	002	1·480	1·475	vw	1·43	w
1·450	vw	112 611	1·449 1·447	1·446	vw	—	—
—	—	550	—	—	—	1·40	(vw)
1·421	vvw	202	1·420	1·415	vvw	—	—
1·394	w	640 541	1·399 1·391	1·390	w–m	1·36 1·35	(w) m
1·370	vvw	222	1·367	—	—	—	—
1·343	vw	312 631	1·343 1·341	1·340	w	—	—
1·327	vw	730	1·325	1·323	vw	1·295	w
1·295	vvw	701	1·296	—	—	—	—
1·277	vw	402	1·277	—	—	1·24	vw
1·257	vw	332 721	1·257 1·255	—	—	1·22	vw
1·238	vw	422	1·237	—	—	—	—
1·190	vw	660	1·189	—	—	—	—
—	—	651 512	—	—	—	1·16	(w)
1·176	vvw	750	1·173	—	—	—	—
1·156	vw	811	1·153	—	—	—	—
1·138	vvw	442	1·139	—	—	—	—
1·116	vw	910	1·114	1·113	w	1·09	(vw)
1·110	vw	602	1·111				
1·099	vw	831	1·103	—	—	—	—
1·086	vvw	622	1·085	—	—	—	—
1·067	vvw	930	1·063	—	—	1·04	(vvw)
1·028	w	712	1·027	1·026	vw	—	—
1·019	vw	770 642	1·019 1·017	—	—	—	—
1·009	vw	10.0.0	1·009	1·007	vw	—	—
0·9897	vw	10.2.0	0·9894	—	—	—	—
0·9627	vw	213	0·9639	—	—	—	—
0·9459	vvw	303	0·9469	—	—	—	—
—	—	10.4.0	—	—	—	0·92	(vvw)
0·9285	vvw	323 662	0·9305 0·9270	—	—	—	—
0·9195	vw	752	0·9193	—	—	—	—
0·9151	vw	413	0·9151	—	—	—	—
0·9133	vvw	11.1.0	0·9135	—	—	—	—
0·8907	vw	880 912	0·8918 0·8902	—	—	0·87	(w)
0·8867	vw	503	0·8864	0·886	w.d.	—	—
0·8730	vw	523	0·8730	0·873	w	—	—
0·8650	w	10·6·0	0·8652	0·864	vw	—	—
0·8479	vw	613	0·8480	—	—	—	—
0·8359	w	543	0·8363	—	—	—	—
0·8298	vw	12.2.0	0·8294	—	—	—	—

Visual estimate of intensity: s strong, m medium, w weak, v very, d diffuse.

[1] Cryptomelane from L. S. Ramsdell (6) in Å. units; lines in parentheses occurred only on a fibre photograph.

Diffraction studies.

X-ray powder diffraction patterns were obtained using two cameras having diameters of 5·73 and 19 cm. The powder specimen was 0·3 mm. in diameter and gum tragacanth was mixed with the powdered priderite to reduce absorption errors. Exposures were made using filtered copper radiation. A list of interplanar spacings and line intensities is given in table II.

Using the method of Hesse (4), indices were assigned to the spacings and the unit-cell dimensions were determined. The dimensions were found accurately by using the 19-cm. camera and measuring the lines of priderite relative to those of sodium chloride; the latter was included in the specimen as an internal standard. Measurements were made in the back reflection region where the α_1 and α_2 lines were resolved.

The unit cell was found to be body-centred tetragonal with the dimensions: a 10·11±0·02 A. (= 10·09 *kX*), c 2·964±0·004 A. (= 2·958 *kX*). This unit cell has since been confirmed by A. D. Wadsley,[1] who has obtained zero and first-level Weissenberg photographs about the a- and c- axes of a crystal.

Discussion.

During this study the close resemblance between the diffraction patterns of priderite and cryptomelane was noticed. For comparison the main lines of cryptomelane from data by Ramsdell (6) are recorded in table II. Cryptomelane has a body-centred monoclinic (pseudo-tetragonal) cell (5–8) with a 9·79, b 2·88, c 9·94 A., β 90° 37′. The formula of cryptomelane approximates to KR_8O_{16} where R is mainly Mn^{4+}.

The c-axes of priderite and rutile are almost the same, while the a-axis of priderite is approximately twice that of rutile. If the density of packing of the oxygen atoms in the two minerals is assumed to be approximately the same (9), then priderite will contain four times as many oxygen atoms as the unit cell of rutile. Table I shows the contents of a unit cell of priderite taking as a basis 16 oxygen atoms. The density of priderite, calculated from the contents and dimensions of a unit cell is 3·948 g./cm.^{3} This is in reasonable agreement with the experimental value of 3·86±0·08 g./cm.^{3}

Because of their similar radii it is assumed, in table I, that the ions Ti^{4+}, Fe^{3+}, and Al^{3+} occupy one set of positions in the unit cell, while

[1] Research officer of the Division of Industrial Chemistry, Commonwealth Scientific and Industrial Research Organization, Melbourne, Victoria.

K^+ and Ba^{2+} occupy another set of positions. The position of the Na^+ ion is doubtful because its radius is between the radii of Ti^{4+} and K^+.

The formula of priderite is approximately $(K,Ba)_{1\cdot33}(Ti,Fe)_8O_{16}$. The ratio of (K+Ba) to (Ti+Fe) is 1:6, an integral relationship which is regarded as purely fortuitous. Since a perfect body-centred unit cell must contain at least two of any one atom, priderite must have a defect structure. Assuming that the structure of priderite is similar to that of cryptomelane its formula may be written as $A_{2-y}B_{8-z}O_{16}$ (7) where A are the large ions K^+ and Ba^{2+} and B are the smaller ions Ti^{4+}, Fe^{3+}, and Al^{3+}; y is of the order of unity, while z is very small.

Priderite has been synthesized by several methods. It was prepared by fusing a powdered mixture of TiO_2 and K_2CO_3 or TiO_2 and $BaCO_3$ in an oxyacetylene flame. The minerals prepared in this way usually contained a small rutile impurity, so that their K_2O and BaO contents were uncertain. However, diffraction patterns of synthetic K-priderite with different K_2O contents were identical with that of the natural mineral. The diffraction pattern of Ba-priderite is recorded in table II and although not identical with that of the natural mineral it indicates that, superficially at least, they are isostructural. The K-priderite was also synthesized by heating to 1000° C. an intimate mixture of KF and TiF_4.

Hollandite and cryptomelane are considered to be members of an isostructural series (5, 7, 9). K-priderite and Ba-priderite appear to be members of an analogous series. Although the structure of priderite has not been determined, all the evidence indicates that it is similar to that of the cryptomelane series (7). A. D. Wadsley is at present carrying out a structure determination on priderite.

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Abstract.—Priderite is the name suggested for a new mineral with the formula $(K,Ba)_{1\cdot3}(Ti,Fe)_8O_{16}$. X-ray diffraction studies show that the

mineral is body-centred tetragonal with a 10·11, c 2·96 Å. Structurally the mineral resembles cryptomelane, KMn_8O_{16}, the Ti of priderite being equivalent to the Mn of cryptomelane Synthetic K- and Ba-priderites appear to be isostructural and they are probably members of a series analogous to cryptomelane–hollandite.

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The crystal structure of some chamosite minerals.

(With Plate XII.)

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With an appendix: *Geological and petrographical notes.*
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[Read in part at the Clay Minerals Group meeting, April 9, 1949.]

1. *Introduction.*

CHAMOSITE is a commonly occurring hydrous ferrous silicate frequently found in close association with chalybite in marine mudstone deposits, the composition of different beds ranging from almost pure chamosite to pure chalybite. In oolitic ironstones chamosite is found both in the ooliths themselves and also in the groundmass. It occurs as a constituent in shales, as a cementing mineral in sandstones, and is occasionally found in lateritic and other deposits. Mineralogically it is usually regarded as a chlorite, or akin to the chlorites, but evidence has been accumulating which tends to show that some varieties may not be chlorites at all, but are probably closely related to the kaolin group of minerals (Brindley, 1949).

Following a preliminary announcement of these results by the present writer at the Clay Minerals Group meeting, April 1949, Orcel, Hénin, and Caillère (1949)[1] examined a number of micaceous minerals in oolitic iron ores both by X-ray and by thermal methods. Out of fifteen specimens they found three with a kaolin- or antigorite-type structure and nine with a chlorite-type structure. Independently, Brindley and Youell have also applied thermal methods and obtained confirmatory evidence that the chamosites previously studied by the writer had indeed a kaolin-type rather than chlorite-type of structure. Commenting on their results, Orcel, Hénin, and Caillère remarked (1949, p. 135): 'On désigne donc sous le nom de chamosite des minéraux cristallographiquement différents.' This clearly indicates that there is a problem of nomenclature to be solved. The writer considers, however, that this question

[1] References, p. 521.

can best be solved by agreement between those who have worked on the structure and identification of this mineral (or these minerals), and in the present paper the term 'chamosite' will continue to be used for minerals which will be shown to have a kaolin-type of structure.

2. *Survey of previous work.*

The earlier classification of chlorites rested largely on chemical analyses (cf. Tschermak, 1891; Orcel, 1927; Hallimond, 1939). Hallimond has discussed the relation of chamosite to the chlorites from the chemical standpoint and has shown how difficult it is to arrive at any firm conclusion. Physical data, however, such as dehydration curves, thermal curves, and more particularly X-ray diagrams, show distinctive differences between chamosites and chlorites. It seems, therefore, not unlikely that differences of structure may exist which are not accompanied by any clearly marked differences of composition.

Hallimond's paper contains a detailed report by Bannister of an X-ray comparison of several chamosites with a number of iron-bearing chlorites. His powder diagrams for six chamosites 'are very similar to each other, but are distinct from those of thuringite, daphnite, &c.' (Hallimond, 1939), which give typical chlorite diagrams. Bannister considered that the chamosite diagrams, though differing among themselves, were sufficiently distinctive to enable this mineral to be recognized from its X-ray diagram, and that the individual differences may have originated from variations in the size and degree of orientation of the crystallites, in their chemical composition, in the unit-cell dimensions, or in their crystal structures. He also noted that other minerals, namely amesite and cronstedtite, at that time regarded as chlorites, gave distinct powder diagrams and both of these have since been shown (amesite by Gruner, 1944, and cronstedtite by Hendricks, 1939) to be closely allied as regards their structures to the kaolin group of minerals.

It was unfortunate that Bannister did not, in this investigation, record the 14 kX line which is very characteristic of chlorite structures. This spacing corresponds to the thickness of one layer of the chlorite structure, and the observation of a 14 kX reflection (if it is not due to an impurity) shows at once that the mineral has essentially a chlorite-type rather than kaolin-type of structure. For reasons given in detail later, it cannot immediately be inferred that a mineral is of kaolin-type if a 14 kX reflection is not observed and a strong line at 7 kX is obtained. In a subsequent investigation (Bannister and Whittard, 1945) of a magnesian

chamosite, so called because of its 'mode of occurrence and physical properties', a strong 14 kX line was observed together with its second, third, and fourth orders. This is typical of chlorite powder diagrams and there are no grounds for doubting that this magnesian chamosite has a chlorite-type of structure. In this paper Bannister also gives X-ray data for an iron-ore chamosite from Frodingham and records the absence of an observable reflection at 14 kX, but (as stated above) this alone provides no certain proof of the nature of the structure.

An investigation of thuringite, bavalite, and chamosite by Engelhardt (1942) showed that the first two are practically identical with each other and closely similar to sheridanite, a typical iron-bearing chlorite; Bannister (see Hallimond, 1939, p. 461) had previously arrived at the same conclusion. Engelhardt also finds that chamosite gives an X-ray diagram different from those of well-authenticated chlorites. He interpreted his chamosite data in terms of a structure closely resembling that of normal chlorites, consisting of layers of the same type, but arranged according to orthorhombic instead of monoclinic axes. Engelhardt's data agree generally with the spacings and intensities to be expected from this structure, but the total number of recorded reflections is not large and the structure cannot therefore be tested very critically.

Engelhardt observed a 14 kX line of weak intensity which is characteristic of chlorites rich in iron. Apart from this line, the evidence for a chlorite structure is rather slender. As regards other basal reflections, the lines indexed as (002) and (004) could be the usual strong lines (001) and (002) of a kaolin-type structure, the existence of (003) is doubtful because it coincides with (020), the (005) line is confused by a chalybite line, and higher orders are not observed. As regards other reflections, nine altogether, their l indices are *even* numbers, which suggests that a cell of half the height may be correct, while their h and k indices correspond to reflections which always tend to be prominent from both chlorite- and kaolin-type structures. The available data, therefore, do not permit a very detailed analysis to be made. Nevertheless, the existence of the 14 kX line points clearly to a chlorite-type of structure and the solution given by Engelhardt is very probably correct for the variety of chamosite which he examined.

Sudo (1943) has given X-ray and other data for a number of low-temperature hydrous silicates found in Japan. He confirms that chamosite has a powder diagram essentially different from that of chlorites and that it is a variable material, but he obtains no further information regarding the structure of this mineral.

3. *The present experiments.*

These have been carried out using chamosites from a variety of sources. They were all microcrystalline materials which could be investigated only by the powder method. In order that the X-ray data should be as complete as possible a 20-cm. diameter semi-focusing type of camera was employed, with filtered Co-$K\alpha$ radiation. The experimental arrangement entails packing the powder into a flat, shallow cavity (about $4 \times 2 \times 0{\cdot}5$ mm.) in a glass plate. When fine powders of 'clay-like' materials are used, no adhesive is required to retain the powder in the holder, and provided the cavity is filled without unnecessary pressure and without smoothing the surface repeatedly, the preferential orientation of the particles is not very marked.

The method requires rather more powder than the minimum necessary to coat a fibre, and in the present experiments the material investigated was carefully selected under a powerful hand-lens and only apparently homogeneous green-coloured material was used. This is not by any means an infallible method for choosing material, but reliance was placed mainly, indeed almost entirely, on the X-ray powder diagrams themselves. From the six most satisfactory materials examined, which came from a wide range of localities, it was clear that the chamosite minerals in all of them yielded very similar patterns of lines (pl. XII). Some differences were noted, which will be discussed later, but they were not of such a kind as to suggest that essentially different minerals were being examined.

Other tests consisted in treating the material with hot, dilute HCl and examining any residue. This served to identify, for example, the presence of several per cent. of anatase and some rutile in one specimen (no. 874, see below) and a large proportion of kaolinite in another specimen, which was therefore discarded. Quartz and any other coarse material present after crushing or light grinding was largely removed by sedimentation.

The writer now places considerable reliance on a combined thermal and X-ray examination of these materials. This will be reported more fully in a separate note by Brindley and Youell, but it may be briefly described here because it has now been applied to three of the chamosites discussed in the present paper and fully confirms the X-ray conclusion, viz. that these chamosites have a structure similar to that of kaolinite. The principle of the method depends on the fact that in a kaolin- or antigorite-type of structure, the (OH)'s are associated with a *single* octahedral layer in the structure, whereas in a chlorite-type of structure

they are associated with two layers, a brucite-type layer and a talc-type layer. In all the cases yet examined, the dehydration of the brucite layer precedes that of the talc layer and the intermediate stage yields a characteristic X-ray pattern (Brindley and Ali, 1950; since this paper was published other chlorites including iron-bearing chlorites have been examined and they give similar results). This does not happen with kaolin-type structures, which appear to dehydrate and decompose in a single process. This method has been used by Orcel, Hénin, and Caillère (1949) and independently by Brindley and Youell.

TABLE I. Chemical analyses of some chamosites.

	No. 874 Ayrshire.	Wabana, Newfoundland.	Frodingham, Lincolnshire.	Schmiedefeld, Thuringia.
SiO_2	22·03	21·40	24·9	26·65
Al_2O_3	22·91	25·40	15·6	16·14
Fe_2O_3	0·46	0·25	7·2	6·69
FeO	36·68	37·60	35·0	34·43
MgO	1·91	2·04	4·6	4·47
CaO	0·07	0·13	—	—
Na_2O	0·08	—	—	—
K_2O	0·03	—	—	—
H_2O+ ...	10·65	12·02	(12·3)	11·42
H_2O- ...	0·63			0·08
TiO_2	3·63	0·89	—	trace
P_2O_5	0·18	—	—	—
MnO	0·04	0·05	0·4	—
CO_2	0·40	0·22	—	—
SO_3	0·27	—	—	—
Cr_2O_3	0·05	—	—	—
Organic ...	0·03	—	—	—
	100·05	100·00	100·00	99·88

No. 874. Geol. Survey Lab. (Analyst, B. E. Dixon.) Data from Mem. Geol. Survey, Scotland, 1930, Economic geol. Ayrshire coalfields, area III.

Wabana, Newfoundland. Analysis and material supplied by Dr. K. W. Andrews, United Steels Ltd., Stocksbridge. Figures given are obtained after deducting free quartz from total SiO_2 and recalculating to 100.

Frodingham, Lincolnshire. Geol. Survey Lab. no. 1015. (Analyst, C. O. Harvey.) After deducting carbonates, &c., and calculating to 100, assuming 12·3% H_2O. (A. F. Hallimond, Min. Mag., 1939, vol. 25, p. 445.)

Schmiedefeld, Thuringia. Data from H. Jung, Chem. Erde, 1931, vol. 6, p. 275. [M.A. 5–40.]

4. *The chamosites used in the present investigation.*

Results will be reported for materials from the following localities:

1. Ayrshire, Scotland. (No. 874, Geol. Surv. Lab.)
2. Knowehead, Co. Antrim, Ireland. (C/1 from Mr. V. A. Eyles, Geol. Surv. Lab.)

3. Carnlough, Co. Antrim, Ireland. (B.0. Mr. V. A. Eyles.)
4. Wabana, Newfoundland. (Wabana, Dr. K. W. Andrews, United Steels Ltd.)
5. Kongsberg, Norway. (B.M. 34121, Dr. F. A. Bannister.)
6. Finedon, Northamptonshire, England. (E. 20568, Prof. J. H. Taylor. Geol. Surv. Lab.)

Geological and petrographical notes on these materials are given in the Appendix (p. 522).

Two only have been chemically analysed and these data are given in table I, together with data for two other chamosites which have been taken from the literature. The analyses of the no. 874 and Wabana materials are very similar to those of other chamosites; they show relatively little Fe_2O_3 and correspondingly more Al_2O_3. The FeO content remains almost constant in these materials.

5. *Powder data for six chamosites.*

Table II records the measured spacings, estimated intensities, and a few details concerning the line profiles of the observed X-ray reflections. The results are contained in columns 1–6; column 7 gives data by Engelhardt (1942) and columns 8 and 9 data by Bannister (Bannister and Whittard, 1945).

The information summarized in this table, together with the powder photographs reproduced in pl. XII, show that, while differences exist between the results for different specimens, there is an overall similarity between them. These minerals have rather imperfect crystalline structures which is indicated by the occurrence of diffuse bands of a type characteristic of irregular layer structures.

The powder diagrams of the present materials differ most remarkably in the group of spacings between 2·7 and 2·3 kX. Specimens 1, 2, and 3 of table II give a trio of clearly defined lines at about 2·67, 2·51, and 2·40 kX; the 2·51 kX line is always the strongest, but the intensities of the two outer lines are variable. Specimens 4, 5, and 6 show the first two lines only, the 2·40 kX line having disappeared entirely, and for specimens 5 and 6 the strong 2·51 kX line has become noticeably weaker and very diffuse. Other lines also are present with specimens 1, 2, and 3, such as the line at 2·01 kX, which are absent from 4, 5, and 6.

The data obtained by Engelhardt and by Bannister, columns 7 and 8 of table II, show general agreement with the data in columns 1 6, especially for the more consistently occurring reflections, but as regards

TABLE II. X-ray spacings and estimated intensities of X-ray powder diagrams of some chamosites.

1 874 Ayrshire.		2 C/1 Co. Antrim.		3 B/0 Co. Antrim.		4 Wabana, Nfld.		5 B.M. 34121 Norway.		6 E. 20568 Finedon.		7 Schmiede- feld.		8 Froding- ham.		9 Wickwar, Glos.	
		—		—		14	½	—		—		14	*	—		14	
7·04	10	7·06	10	7·05	10	7·07	10	7·15	9	7·17	9	6·93	10	7·1	10	7·1	10
4·66	2†	4·66	3	4·66	3†	4·67	4†	4·61	3†	4·65	2†	4·65	5	4·63	5	4·73	2; 4·69 7
4·57	2	4·56	3	4·59	3	—		—		4·58	½	—		—		—	
4·27	½	g		g		—		—		—		—		—		—	
3·891	1	3·882	3	3·888	2	3·89	½	3·898	½	—		—		—		—	
3·513	10	3·513	10	3·514	10	3·524	10	3·553	9	3·547	9	3·50	9	3·53	9	3·53	10
2·796	?	—		—		—		—		—		2·779	3 s	—		2·83	5
2·673	4	2·667	8	2·668	9	2·699	3	2·675	2	2·681	4	2·689	1	2·69	7	2·69	2; 2·61 2
2·514	9	2·510	10	2·508	10	2·516	7	2·574	5†	2·516	6†	2·507	7	2·53	7	2·52	1; 2·60 1
2·399	4	2·396	9	2·397	10			—		—		—		2·47	4	2·46	4
														2·40	4	2·39	4
2·332	½	2·346	1	2·340	½	2·342	½	—		2·330	½	2·330	½	—		—	
2·269	½	2·274	1	2·269	½	2·274	½	2·28	?	2·28	?	—		—		2·28	2
2·137	6	2·135	5	2·134	4	2·143	4 b	2·143	½	2·140	1	2·134	3	2·16	2	2·14	1
2·007	1	2·007	5	2·006	3	—		—		1·991	½	—		2·02	1	2·08	1; 2·01 5
1·890	1	1·888	½	1·883	½	—		—		1·88	?	—		—		1·89	2; 1·83 1
1·765	4	1·764	3	1·764	2	1·765	2 b	1·778	½ b	–		1·767	2	1·78	1	—	
1·69	? a	—		—		—		—		—		—		1·74	1	—	
														1·72	1	1·72	2
1·662	½ a	1·661	1	1·659	1	1·656	?	1·66	?	1·65	?	—		1·68	2	1·67	2; 1·63 1
																1·573	3
1·552	7	1·553	7	1·550	8	1·558	6	1·549	2 b	1·554	5	1·556	5	1·557	8	1·552	6
1·518	3	1·519	3	1·515	2	1·522	3	1·515	½	1·518	2	1·524	2	1·524	7	1·517	4
		1·484	2 b	1·488	1 b			1·484	½	1·49	?					1·479	1
1·470	1	1·469	1	—		1·477	½	—		—		1·473	1	—		—	
1·422	1	—		-		1·422	½	1·42	?	1·421	½	1·425	1	1·427	4	1·424	4
1·404	½	—		—		—		—		-		—		—		1·395	4
1·358	½	—		-		—		—		—		—		1·350	2	1·361	1
1·344	½	—		—		-		—		-		-		—		—	
1·323	½	—		—		—		1·32	?	—		-		1·328	2	1·333	1

(1) to (6), Brindley and Goodyear (not previously published). Filtered Co-K_α radiation; camera radius, 100 mm.

(7), by v. Engelhardt (1942). Filtered Fe-K_α radiation, camera radius 57·5 mm.

(8), (9), Bannister and Whittard (1945). Unfiltered Fe—K radiation; camera radius 30 mm.

So far as is known specimens (1) to (8) are 'normal' chamosites; (9) is a magnesian chamosite.

a, lines wholly or partly due to anatase; b, unusually broad line; g, additional lines in this region probably due to goethite impurity; s, doubtful chalybite line; *, observed, but not measured; †, the head of a diffuse band shading to smaller spacings.

the 2·67, 2·51, 2·40 kX trio, Engelhardt finds only two lines and his material appears to resemble most closely the present no. 4. Bannister's data for the magnesian chamosite, column 9, contain many more lines than are ever found for 'normal' chamosites. Bannister's earlier data (Hallimond, 1939) also show general concurrence, but he gave consistently smaller spacings than those recorded in table II, especially for the lower orders.

The importance of determining whether a reflection occurs from a spacing of about 14 kX has already been stressed and has been studied with special care. No reflection which can be observed in powder diagrams occurs in this region from specimens 1, 2, 3, 5, and 6. Specimen no. 4 gave a very weak line at about 14 kX, which is probably due to a trace of chlorite impurity: this is also suggested by the thermal investigations by the writer and Youell (see section 11).

6. *Has chamosite a chlorite type of structure?*

This question requires mainly a consideration of the intensities of the basal reflections. Chamosite does not give very well ordered aggregates, but specimens have been produced sufficiently well orientated to show that the strong reflections at 7·05 and 3·51 kX are basal reflections and from these the higher orders can be calculated and compared with recorded values. If the structure is of chlorite type, and if the thickness of a single chlorite layer, about 14 kX, is accepted as the basal spacing, then the 7·05 and 3·51 kX lines will be respectively (002) and (004). In what follows the data for specimen no. 1 (no. 874 from Ayrshire) will be mainly considered, since this was the only material for which a detailed chemical analysis was available when this work was begun. It also gives the clearest X-ray photograph. The analysis of this material (see table I) is very similar to that of other chamosites, especially as regards the iron content, and the calculation of the reflected X-ray intensities would be modified very slightly by using other chamosite analyses. The intensities depend very much more on the type of structure which is assumed than on *small* changes of composition.

In the ideal chlorite structure, the brucite layer has the composition $Mg_6(OH)_{12}$, and the talc or mica layer $Mg_6(Si_8O_{20})(OH)_4$: the combined formula is $Mg_{12}(Si_8O_{20})(OH)_{16}$. The analysis in table I for the no. 874 material leads to the following chlorite-type formula:

$$\underbrace{(Al_{3\cdot05},Fe^{3}_{0\cdot08},Cr_{0\cdot01},Fe^{2+}_{6\cdot90},Mg_{0\cdot64},Ti_{0\cdot61})}_{11\cdot29\text{ atoms}}\ (Si_{1\cdot27},Al_{3\cdot03})O_{20\cdot0}(OH)_{16\cdot0}$$

It is doubtful if the TiO_2 recorded in the analysis should be included in the chamosite composition, since X-ray examination reveals anatase and possibly rutile after dissolving out the chamosite with warm HCl. Since it is not certain that all the TiO_2 can thus be accounted for, the analytical figures have been accepted without corrections; in any case it makes little difference to the formula. It is satisfactory to find that the O/OH ratio works out to exactly 20 : 16. Engelhardt's formula for the material which he examined is as follows:

$$\underbrace{(Al_{2\cdot6},Fe^{3+}_{1\cdot2},Fe^{2+}_{6\cdot7},Mg_{1\cdot5})}_{12\cdot0\text{ atoms}}\ (Si_{6\cdot2},Al_{1\cdot8})O_{20\cdot0}(OH)_{17\cdot7}$$

The main difference lies in the extent to which Si is replaced by Al; the 'octahedral atoms' are roughly the same.

TABLE III. Comparison of F^2_{001} calculated for chlorite-type structures and observed data.

Structure (*a*), Fe atoms filling all sites in brucite layer.
Structure (*b*), Fe atoms randomly distributed.
Structure (*c*), Fe atoms filling all sites in talc layer.

	Calculated data				Observed data		
		$F^2 \times 10^{-2}$					
hkl.	*d* (kX).	(*a*).	(*b*).	(*c*).	*d* (kX).	Intensity.	Notes.
001	14·05	2	25	128	(14)	usually absent	1
002	7·03	330	330	330	7·04–7·06	10	—
003	4·684	333	156	47	4·65–4·67	2–3	2
004	3·513	487	487	487	3·513–3·514	10	—
005	2·810	32	109	230	—	—	—
006	2·342	27	27	27	2·330–2·346	0·5	—
007	2·007	108	41	5	2·006–2·007	3	—
008	1·756	5	5	5	—	—	—
009	1·561	2	4	30	—	—	3
0010	1·405	187	187	187	1·404	0·5	4
0011	1·277	0	10	40	—	—	—

Notes:

1. A very weak line is observed for specimen 4 only, which probably arises from an impurity.
2. The observed spacing is always less than (003). The reflection shows no enhancement due to orientation and its diffuseness and spacing both suggest (020) rather than (003).
3. An observed strong line at 1·550 1·558 is much more likely to be (060) than (009). Note that 4·65, the spacing of the probable (020), is exactly $3 \times 1{\cdot}55$.
4. This line was only recorded from an orientated specimen of no. 1.

The distribution of the 'octahedral atoms' between the brucite and talc layers cannot be decided immediately. Structures have been considered in which iron atoms occupy (*a*) all available positions in the

brucite layer, (*c*) all available positions in the talc layer, and (*b*) positions randomly distributed throughout both layers. The F^2_{001} values calculated for these three distributions are tabulated in table III, where they are compared with the spacings and intensities of the nearest observed reflections.

It is clear from table III that the observed data do not agree even qualitatively with any one of the three distributions. Moreover, since (*a*) and (*c*) are extreme cases in which the iron atoms are placed so far as possible wholly in the brucite or in the talc layer and (*b*) is an average case, one cannot hope to find any other distribution of the iron atoms in a chlorite-type structure which will fit the observed data more satisfactorily. Minor changes in composition would not materially alter this conclusion. *It is on these grounds that the conclusion is drawn that the chamosites examined in this investigation are not chlorite-type minerals.*

One result given in table III calls for special comment, namely the very weak intensity calculated for the 14 kX reflection when the iron atoms are concentrated in the brucite layer, i.e. arrangement (*a*). It is clearly possible for a chlorite-type structure to exist without any observable 14 kX line, but (and this is very important) there must also be strong or medium-strong (003) and (007) reflections. Engelhardt chose arrangement (*b*) for his structure and remarked that neither (*a*) nor (*c*) would fit the observed facts. But for the chamosites examined in the present work, (*b*) also is obviously wrong, since no trace of a line is found to correspond to (005) at 2·81 kX. It is unfortunate that Engelhardt's data are not conclusive on this point, since he suspected a chalybite line in this position. Moreover, the medium-strong (003) of arrangement (*b*) does not really fit the observations, since the observed line, although *near* to the (003) position, is certainly not correctly spaced for (003) and the character of the line, its band-like appearance, suggests (020) by comparison with other layer minerals. *The conclusion seems quite unavoidable that the structure is not that of a chlorite.*

Two other layer structures have been considered giving a 14 kX line and its higher orders, both of which were also considered by McMurchy (1934) in his work on the structure of chlorites. The first consists of kaolin-type layers with the kaolin units pointing alternately in opposite directions along the *c*-axis. The second arrangement consisted of the following sequence of layers: a kaolin layer, a talc layer, a kaolin layer of opposite orientation to the first, and a brucite layer. Both arrangements predict results quite incompatible with the observed data.

7. *Has chamosite a kaolin-type structure?*

The ideal formula of kaolinite is $Al_2(Si_2O_5)(OH)_4$, and of antigorite, $Mg_3(Si_2O_5)(OH)_4$. The composition of the chamosite specimen, no. 874, can be similarly written, namely:

$$\underbrace{(Al_{0\cdot76}, Fe^{3}_{0\cdot02}, Fe^{2}_{1\cdot73}, Mg_{0\cdot16}, Ti_{0\cdot15})}_{2\cdot82\text{ atoms}}(Si_{1\cdot24}, Al_{0\cdot76})O_5(OH)_4.$$

There are therefore 2·82 atoms to occupy three octahedral positions, so that the composition is more nearly related to antigorite than to kaolinite.[1]

Preliminary calculations of the basal intensities from a kaolin-type structure employing the atomic separations found in kaolinite itself showed very promising results; a more careful consideration was therefore given to the effect on the interatomic distances of replacing about 3 out of every 8 silicon atoms by aluminium and of filling octahedral positions with the larger Fe^{2+}ions.

The size of the tetrahedral Si(Al)–O group may be estimated from the basal plane dimensions. In kaolinite (Brindley and Robinson, 1946) b is 8·93 kX and in chamosite (see later) b is 9·31 kX. If the tetrahedral group is correspondingly expanded, the Si O separation of 1·60 kX in kaolinite becomes 1·67 kX in chamosite, and the height of the tetrahedron, 2·19 kX in kaolinite, becomes 2·28 kX in chamosite. The value finally adopted for the height of the tetrahedron in chamosite is 2·34 kX, which gives a mean (Si,Al)–O distance of 1·71 kX.

The dimensions of the octahedral layer can be estimated from the known radius of the Fe^{2+} ion, which is about 0·83 A. as compared with 0·57 A. for Al. Assuming a regular octahedral grouping and taking the radius of oxygen to be 1·32 A., we find for the thickness of an octahedral layer containing Al, 2·18 kX, and containing Fe^{2+}, 2·48 kX. Since the thickness found in kaolinite, 2·12 kX, is rather less than the calculated value, the same may be true for chamosite. The value finally adopted (arrangement (iii) of table IV) is 2·36 kX.

The observed and calculated reflected intensities are set out in table IV. The observed values include three reflections at 1·404, 1·165, and 1·004 kX obtained only with an orientated specimen of chamosite no. 874, which are not included in table II. Under the heading 'calculated data' are

[1] I shall, however, continue to refer to the structure as being of 'kaolin-type', meaning thereby that it is built of a tetrahedral Si–O sheet combined with an octahedral X O(OH) sheet, which is characteristic of all the kaolin minerals. In this sense, the *structure* (though not the *composition*) of antigorite is also of kaolin-type.

given values of $F^2A\phi(\theta)$ where F is the structure factor of the cell, A the absorption factor, and $\phi(\theta)$ includes the polarization and other angular factors which occur in the full expression for the reflected intensity, and θ is the Bragg angle of reflection. These are to be compared with the observed intensities in the final column. The table shows values for four sets of atomic parameters; altogether nine have been considered. While these four sets of parameters reproduce the main features of the experimental results, namely very strong (001) and (002) reflections and the remainder weak or absent, there is a progressive improvement in going from (i) to (iii) or (iv). The observed (001) and (002) reflections have comparable intensities with (002) just greater than (001) which agrees with arrangement (iv). The fact that (003) and (004) are absent is explained best by arrangement (iii). The last three reflections are scarcely visible and the fact that the calculated values exceed very little, if at all, the values for (003) and (004) is probably about right. On the whole, arrangement (iii) gives the best general agreement with the observed data. It is interesting to note that with these parameters, the layers of oxygen atoms are equally spaced throughout the structure, their heights above the basal plane being 0, 2·34, 4·70, and 7·03 kX.

TABLE IV. Comparison of observed and calculated basal reflections from chamosite assuming various kaolin-type structures.

I. Heights of atoms above basal plane of unit cell, in kX.

	(i).	(ii).	(iii).	(iv).	Kaolinite.
oxygens	0	0	0	0	0
48 silicons, 1·52 aluminiums ...	0·66	0·66	0·66	0·66	0·60
oxygens, 2 hydroxyls. ...	2·34	2·34	2·34	2·34	2·19
64 'octahedral atoms', Fe^{2+}, Al	3·33	3·42	3·52	3·58	3·25
hydroxyls	4·34	4·50	4·70	4·82	4·31

II. Comparison of observed and calculated basal reflections.

	Calculated data					Observed data	
		$F^2A\phi(\theta)\times10^{-3}$					Estimated
hkl	*d* (kX).	i.	ii.	iii.	iv.	*d* (kX).	intensity.
001	7·026	448	367	276	219	7·04–7·06	10
002	3·513	329	315	268	236	3·513–3·514	10
003	2·341	19	14	11	11	2·333–2·346	0·5
004	1·756	3	6	13	21	—	—
005	1·405	13	16	23	27	1·404	0·5
006	1·170	17	18	16½	14	1·165[1]	0·5
007	1·004	9	11	12	18	1·004[1]	0·5

[1] These reflections were only recorded from an orientated specimen of no. 874 and are not included in table II.

The agreement between the observed and calculated values is obviously much better for the kaolin-type than for the chlorite-type structure. The final decision, however, must be based on a full consideration of the whole diffraction pattern. The agreement shown in table IV is, however, amply good enough to justify carrying the analysis further.

8. *A kaolin-type unit cell for chamosite.*

The unit-cell dimensions may be determined in a manner similar to that used for other layer silicate minerals. These generally give a strong (060) reflection in the neighbourhood of 1·5 kX, and the strong line consistently obtained at about 1·55 kX for chamosite may be tentatively indexed as (060). This is justified later (cf. table V). A comparison with the kaolin family of minerals suggests that the line at about 4·6 kX may be the (020). Considering the numerical data for specimen no. 874 of table II, we then have $d_{(020)}$ 4·66 kX, b 9·32 kX; $d_{(060)}$ 1·552 kX, b 9·312 kX.

For layer silicate minerals a is approximately $b/\sqrt{3}$. We therefore have provisionally a 5·38, b 9·31 kX, and the thickness of the layer can be taken as $2 \times d_{(002)} = 7{\cdot}026$ kX. It remains to consider the direction of the c-axis. If the cell is monoclinic, it remains only to find the β-angle and this can be done by comparing $(1/d)^2$ for the observed reflections with a calculated plot of $(1/d)^2$ against β for all possible reflections.

This procedure leads to two values of β, each of which accounts for the majority of the reflections, namely, β 90° and β 103° 104°. The first value gives an ortho-hexagonal cell which recalls Engelhardt's solution, while the second gives a monoclinic cell closely similar to that of kaolinite, though the latter is strictly triclinic, as shown by Brindley and Robinson (1946). Although many lines are common to both these unit cells, at least so far as their *spacings* are concerned, certain lines, including some of quite considerable intensity, can be explained only by the one or the other cell. This raises the question whether the true unit-cell is larger than either and contains both of them as sub-units, or whether actual chamosites are mixtures of two forms.

The first of these possibilities was extensively investigated. The β-angle of about 104° corresponds to a displacement of successive layers by $a/3$, a displacement which occurs in many layer silicate minerals. As previously described (Brindley, 1949), an orthorhombic cell of height about 3×7 kX containing three kaolin-type layers is a possible solution so far as purely geometrical questions are involved (text-fig. 1): all the observed reflections can be indexed with this type of cell, and, if correct,

this would be a very interesting solution, because Hendricks (1939) has already arrived at a similar conclusion for cronstedtite. The immediate objections to it are: (*a*) that it predicts a much larger total number of reflections than are observed, and (*b*) that a 3-layer cell suggests a well-organized crystalline structure which, on general grounds, seems unlikely for a material so poorly crystalline as chamosite. Nevertheless, many arrangements of kaolin-type layers in a 3-layer cell have been systematically examined, but in every case they predicted strong reflections in positions where none are observed.

Further consideration was therefore given to the second possibility, namely, that two forms of chamosite exist, a monoclinic and an orthorhombic or ortho-hexagonal form. It was considered that such a mixture might explain the marked variations observed in chamosite powder diagrams and, as will now be shown, this seems to be confirmed by the detailed results.

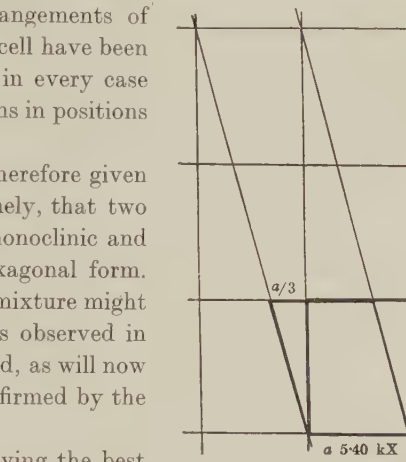

FIG. 1. Comparison of ortho-hexagonal and monoclinic unit-cells. The two cells are shown by heavy lines in projection on the *ac*-plane.

The unit-cell dimensions giving the best overall agreement with the measured spacings are:

Ortho-hexagonal cell: *a* 5·38, *b* 9·314, *c* 7·026 kX.

Monoclinic cell: *a* 5·40, *b* 9·314, *c* sin β 7·026 kX, β 104·5°.

The *a* parameter of the ortho-hexagonal cell was taken to be exactly $b/\sqrt{3}$ = 5·38 kX which differs scarcely appreciably from the figure 5·40 kX chosen for the monoclinic cell; no real difference is implied between these two parameters. The symmetry of the atomic arrangement is $C3m$, provided the octahedral positions are occupied randomly by Fe^{2+}, Mg, Al . . . atoms. For the monoclinic cell, *a* 5·40 was found to give the best agreement with the measured spacings and in so far as all the octahedral positions are equally filled, the space-group is Cm.

9. *Comparison of observed and calculated data.*

The calculated spacings and intensities for the hexagonal and monoclinic structures are set out in table V together with the observed data for specimen 1 of table II. The spacings have been calculated down to

TABLE V. Comparison of observed data with calculated data for monoclinic and hexagonal structures.

I (calc.) $= pF^2\phi(\theta)\times 10^{-4}$.
I (obs.) = Visually estimated intensities on scale 10, 9, 8 . . . 1, ½.

Monoclinic structure			Specimen no. 874 (See table II)		Hexagonal structure		
d(kX).	hkl.	I (calc.)	d(kX).	I (obs.)	d(kX).	hkl.	I (calc.)
7·026	001	70	7·04	10	7·026	001	70
4·657	020	7	4·66	2	4·657	020	21
4·559	110	14	4·57	2			
4·274	11$\bar{1}$	11	4·27	½			
3·880	021	8	3·891	1	3·880	021	25
3·513	002	38	3·513	10	3·513	002	38
3·492	111	6					
3·132	11$\bar{2}$	4					
2·804	022	3	2·796	?	2·804	022	7
2·679	20$\bar{1}$	13	2·673	4	2·688	200	8
2·669	130	19					
			2·514	9	2·511	201	88
2·405	20$\bar{2}$	27	2·399	4			
2·396	131	55					
			2·332	½			
2·271	201	4					
2·270	13$\bar{2}$	9	2·269	½			
					2·210	041	2
			2·137	6	2·135	202	48
2·012	20$\bar{3}$	8					
2·005	132	15	2·007	1			
					1·941	042	3
1·898	13$\bar{3}$	5	1·890	1			
1·884	202	3					
			1·765	4	1·766	203	30
					1·760	240	2
			1·69	?	1·707	241	3
1·664	20$\bar{4}$	6	1·662	½			
1·658	133	12			1·651	043	2
1·562	13$\bar{4}$	13					
1·561	203	5					
1·558	33$\bar{1}$	14	1·552	7			
1·552	060	7			1·552	060	21
1·521	33$\bar{2}$	3					
1·520	330	5	1·518	3			
1·516	061	4			1·516	061	11
			1·470	1	1·471	204	23
1·425	33$\bar{3}$	2					
1·422	331	2	1·422	1			
1·420	062	2			1·420	062	6
1·405	005	3	1·404	½[1]	1·405	005	3
					1·403	044	2
			1·358	½			
1·347	40$\bar{1}$	1					
1·343	26$\bar{1}$	5	1·344	½	1·344	400	2
			1·323	½	1·320	401	14
1·314	13$\bar{5}$	5					
1·314	204	3					
1·307	400	3					
1·303	26$\bar{2}$	4					

[1] Obtained from ordered aggregate specimen only.

a value of 1·3 kX and since both cells are centred on the *c*-face, reflections occur only when $(h + k)$ is an even number. The calculated intensities have been taken to be given by the expression $pF^2\phi(\theta)$, where F is the structure factor of the cell, p the multiplicity factor, and

$$\phi(\theta) = (1+\cos^2 2\theta)/(\sin^2\theta\cos\theta),$$

where θ is the Bragg reflection angle. No allowance has been made for absorption, which if included would reduce the calculated values for the lower orders relative to those for the higher orders. The total number of possible reflections from spacings greater than 1·3 kX is very great, being 88 for the monoclinic structure and 30 for the hexagonal. Many of these are extremely weak, and to avoid giving an unnecessarily extensive table, the weakest reflections have been omitted from table V. The strongest reflections have values of $pF^2\phi(\theta)$ of the order of 100×10^4 and all reflections weaker than $2 \cdot 10^4$ are omitted. In general, only reflections with $k = 3n$, where n is integral, are sufficiently strong to be incorporated in the table.

To facilitate the comparison of the observed and calculated results, the observed values for specimen 1 alone are incorporated in table V. It is evident that certain observed reflections are wholly monoclinic in origin, others are wholly hexagonal, while others are of mixed origin. It is seen that all the strong and medium-strong lines calculated for the two structures are present in the observed data with about the right intensities if the two types of structure are present in roughly equal proportions. Furthermore, all the observed spacings can be correlated very closely with the calculated values. The calculated results are compared with the powder diagram of material no. 874 in pl. XII.

It seems worth while, therefore, to inquire a little more carefully into the question of the relative proportions of the two components. If the important trio of lines already mentioned, at 2·67, 2·51, 2·40 kX, is considered, it is seen that a mixture of the two components of varying proportions will explain at least qualitatively the marked variation which is observed in their relative intensities. It will be sufficient to examine quantitatively the group of reflections from 2·67 down to 1·52 kX; higher-order lines are too weak to be usefully considered and likewise the lower orders, apart from the basal reflections which are of no help in this connexion.

Table VI sets out a comparison between the observed and calculated relative intensities for different proportions of the two components (monoclinic and hexagonal) for the chamosite specimens 1–6 considered in table II. The observed intensities are visual estimates and therefore

should not be taken too precisely; the calculated intensities are the sums of the $pF^2\phi(\theta)$ values scaled in each case to the value given for the 2·51 kX line. The assumed percentages of the two components are given at the head of each column. It is seen that the scheme accounts reasonably well for the observed variations of intensity from specimen to specimen and especially for the marked changes of the 2·67, 2·51, 2·40 kX trio.

TABLE VI. To illustrate the extent to which the powder data of chamosite specimens 1–6 can be interpreted in terms of a mixture of monoclinic (M) and hexagonal (H) forms.

	55 % M, 45 % H			33 % M, 67 % H		10 % M, 90 % H			
d (kX) Observed, Specimen (1)	I (Calc.)	I (Obs.) Specimens (2)	(3)	I (Calc.)	I (Obs.) Specimen (1)	I (Calc.)	I (Obs.) Specimens (4)	(5)	(6)
2·673	5	8	9	2½	4	1	3	2	4
2·514	*10*	10	10	*10*	9	7	7	5	6
2·399	11	9	10	4½	4	½	0	0	0
2·269	2	1	½	½	½	0	½	0	0
2·137	5½	5	4	5½	6	4	4	½	1
2·007	3	5	3	1	1	0	0	0	½
1·890	1	½	½	½	1	0	0	0	?
1·765	3½	3	2	3½	4	2½	2	½	0
1·552	7½	7	8	4½	7	2	6	2	5
1·518	3	3	2	2	3	1	3	½	2

10. *Disorder in the chamosite structures.*

Fine-grained hydro-silicate minerals having structures of a layer type, such as the clay minerals, commonly show diffraction effects due to irregular stacking of the layers. A very frequent type of irregularity arises from displacements of layers parallel to the b-axis by integral multiples of $b/3$ and if these displacements are strictly random, then (hkl) reflections occur only when $k = 3n$. Now in the case of chamosite almost all reflections with $k \neq 3n$ are in any case too weak to be observed; note that in table V all reflections are recorded for which $pF^2\phi(\theta) > 2 \times 10^4$ and practically the only reflections with $k \neq 3n$ which can be observed separately are the (020), (110), and (11$\bar{1}$). Of these, the (11$\bar{1}$) is extremely weak and the (020) and (110) occur as a very close doublet. These reflections, however, show the characteristic banded appearance found with other minerals in which the layers are randomly displaced. Such displacements therefore probably exist in chamosite as in many other layer silicates.

An important question is whether the two forms (hexagonal and

monoclinic) occur separately or are closely intergrown. Since they differ only in the way the layers are stacked, there is no geometrical reason why they should not be intergrown. The question arises whether the two forms may occur in the form of a single structure with 'mistakes' in the sense used by Wilson (1949) in describing the structure of cobalt; this metal normally consists of a mixture of cubic and hexagonal forms as a result of 'mistakes' in the stacking sequence. The monoclinic and hexagonal forms of chamosite differ only in the arrangement of the layers relative to the a-axis, the monoclinic structure having an $a/3$ displacement and the hexagonal structure zero displacement in the a-direction between layers. A structure which contained a statistical distribution of a 3 and zero displacements between layers would give (hkl) reflections only for h = 0, 3, 6, ... Actually many strong reflections occur with h = 1 and 2. The two forms of chamosite cannot therefore be attributed to random a-axis displacements of layers within a single structure.

If the two forms are intergrown, the individual units must therefore be sufficiently well developed to diffract as individual crystals. Alternatively, the two forms may exist separately. Attempts to modify the ratio of the two forms by a differential acid attack has not yielded positive results.

11. *Thermal study of chamosite.*

Through the courtesy of Prof. A. L. Roberts of the Fuel Department of this University, Dr. R. W. Grimshaw recorded differential thermal analysis curves of specimens no. 874 and B/O. Since about half a gram of material is required for this purpose it is more difficult to obtain this quantity free from impurities than the small quantity required for X-ray analysis. The sample of B/O contained some impurities, mainly goethite and gibbsite, which also contributed to the thermal record. The goethite was also found in the X-ray examination, while the gibbsite was also recognized by X-rays, but was avoided in the material chosen for X-ray examination. The main fact which emerges from the thermal analysis is that the chamosite in both specimens shows a marked endothermic peak commencing at about 450° C. and having a maximum at about 550° C. This is very similar to the thermal behaviour of kaolinite, halloysite, and kaolin-type fireclays (Grimshaw, Heaton, and Roberts, 1945).

Following the X-ray and thermal study of magnesian chlorites by Brindley and Ali (1950), Youell and the writer have applied a similar

analysis to daphnite (an iron chlorite of very similar chemical composition to chamosite) and to the following chamosite specimens: no. 874 from Ayrshire, B/O from Co. Antrim, and the Wabana material. The enhancement of the 14 kX line from daphnite due to dehydration of the 'brucite' layer of the structure was entirely in accordance with the behaviour of other chlorites. The chamosites, however, gave no such effect, but showed a decomposition resulting in an amorphous state presumably similar to that of the kaolin minerals when heated to about 550° C. The very weak 14 kX line shown by the Wabana material was considered with particular care; when the bulk of the material had decomposed, this weak 14 kX line remained slightly enhanced. These results clearly suggested that the bulk of this material was similar to the other chamosites and of kaolin-type, with a small amount of chlorite impurity.

12. *Summary and conclusions.*

Previous X-ray examinations of chamosites by Bannister and by Engelhardt are discussed and also the recent X-ray and thermal studies of Orcel, Hénin, and Caillère on phyllitic minerals in iron ores. Engelhardt suggested a chlorite-type structure for a Schmiedefeld chamosite, while Orcel et al. find evidence for both chlorite-type and kaolin- (or antigorite-) type structures.

X-ray powder data are presented for six chamosites from different localities. Detailed consideration of the recorded reflections and their intensities shows that a chlorite-type structure is unlikely for these materials. A kaolin-type structure is shown to fit the observed data, but two structural forms are thought to co-exist, a monoclinic form and an ortho-hexagonal form. The ratio of monoclinic to ortho-hexagonal form ranges from about 55/45 to 10/90.

For one analysed chamosite form Ayrshire, the formula can be written

$$(Al_{0\cdot76},Fe^{3}_{0\cdot02},Fe^{2}_{1\cdot73},Mg_{0\cdot16},Ti_{0\cdot15})(Si_{1\cdot24},Al_{0\cdot76})O_5(OH)_4.$$

(2·82 'octahedral' atoms)

Thermal examination of three of the six materials is briefly described. Differential thermal analyses by Grimshaw and combined thermal and X-ray study by Brindley and Youell support the view that these minerals have kaolin-type structures.

Finally, there is a question of nomenclature to be considered. The name chamosite is now so widely used and understood that probably only confusion would result if an attempt were made to restrict its use to either a chlorite-type or a kaolin-type of mineral. On the other hand,

if (as now seems certain) both structural types exist, then the question arises whether one name can continue to be used for two structural types. The writer prefers to leave this question of nomenclature open for the present.

Acknowledgements.—Finally, I wish to record my thanks to Dr. K. W. Andrews (United Steels Ltd., Stocksbridge), Dr. F. A. Bannister (British Museum), Mr. V. A. Eyles (Geological Survey), and Prof. J. H. Taylor (King's College, University of London) for materials used in this investigation, and for discussions on the chamosite problem. I wish also to thank Prof. K. C. Dunham (Durham University), Mr. Eyles, and Prof. Taylor for contributing the geological and petrographical notes given in the Appendix: it should be stated that this does not imply their acceptance of the views expressed in the paper.

I wish to acknowledge also that the data recorded in table II were obtained partly by Dr. J. Goodyear when working with me on the mineralogical composition of certain lateritic deposits. In connexion with the thermal study of chamosites, I wish to thank Dr. R. W. Grimshaw for differential thermal curves and Mr. R. F. Youell for assistance with the combined X-ray and thermal examination.

To the Department of Scientific and Industrial Research and to the Government Grant Committee of the Royal Society I tender thanks for grants providing, respectively, a research assistant and equipment.

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APPENDIX

Geological and petrographical notes on the chamosite materials employed in this investigation.

By Prof. K. C. Dunham, Mr. V. A. Eyles, and Prof. J. H. Taylor.

Geological data for specimens 874, C/1, and B/0. By V. A. Eyles.

Material from Ayrshire (Geological Survey Lab. nos. 754 and 874).—Hard greenish-grey clay of Millstone Grit age, occurring in association with other clays of varying composition, some of which are lateritic in origin and derived from the alteration of olivine-basalts.

Material from Knowehead, Co. Antrim. (C/1).—Green-stained basaltic lithomarge (a kaolinized basalt) occurring in association with the Tertiary basalts of Antrim. The production of chamosite in the rock is believed to result from the action of reducing solutions acting on ferric oxide (or hydrated oxide) and halloysite present in the lithomarge.

Material from Carnlough, Co. Antrim. (B/0). Green and buff lateritic lithomarge occurring in association with the Tertiary basalts of Antrim; a heterogeneous rock, the green portions of which contain chamosite probably similar in origin to the Knowehead chamosite (C/1), other parts of the rock, containing gibbsite, being truly lateritic in origin.

It is probable that the Ayrshire chamosite was formed under similar conditions to that from Knowehead and Carnlough, though the field evidence is not so clear on this point. I should think, in the case of the two specimens from Ireland, the chamosite was formed from halloysite;

nd the Ayrshire chamosite could well have been formed in the same vay, though possibly from kaolinite rather than halloysite. In each ocality the mineral was formed, not by sub-aerial weathering, but as he result of climatic conditions at the surface which caused the forma- ion and downward percolation of *reducing* solutions (laterite and litho- narge were originally formed under *oxidizing* conditions).

So far as I have investigated the matter, I do not think, either in reland or Ayrshire, that the chamosite, once formed, has suffered any lteration from sub-aerial weathering processes.

In the case of the Irish occurrences, it can be said that the rocks ontaining chamosite probably suffered a very mild degree of baking as result of the heat action of later flows of basalt that covered up the nterbasaltic horizon, and though the temperature attained is likely to ave been somewhat lower than 150° C. it may have been maintained or a considerable time. As to pressure, the rocks were, subsequently to he formation of the chamosite, buried under a cover of some hundreds f feet of basalt.

In the case of the Ayrshire (Millstone Grit) chamosite, the rocks in hich it is contained must have been buried subsequently to a depth f the order of 4000 feet by later formations, mainly sediments -in fact, have suggested elsewhere that the formation of boehmite and some iaspore in associated rocks results from the pressure of the super- ncumbent rocks. Some rise in temperature as a result of burial is also mplied of the order of 30° to 40° C. above surface temperature, pos- ibly rather more than this, as later the rocks were pierced by dikes, sills, nd volcanic vents.

Petrographical data for specimens 874, C/1 and B 0. By K. C. Dunham.

Material from Ayrshire. (Geol. Survey Lab. no. 874; slice no. S. 22084.) 00 yards W. 20° N. of Gateside farm, 2 miles E. of Ayr; 1-inch S. 14, -inch Ayr 33 NE.—A pale greenish-grey rock, containing what appear o be angular fragments up to 0·15 mm. long of clear, pale-green, feebly- irefringent material. These are set in a matrix of very pale-green aterial, cloudy with tiny inclusions of the order of a few microns in iameter. Scattered through the rock there are opaque granules, mostly hite by reflected light (opal or anatase?). The pale-green material has efractive indices within the range 1·640–1·650. A few small areas of halybite, marginally oxidized, are present. (Anatase confirmed by -rays, G.W.B.)

Material from Knowehead. (V. A. Eyles, C/1, Geol. Survey slice

no. 1.988.) 400 yards WNW. of Knowehead farm, 3 miles S. by W. of Ballyclare: 1-inch I. 28, 6-inch Antrim 51 SE.—Lithomarge preserving the texture of a basalt, but in a highly altered condition. The felspars have been wholly replaced by a colourless or very faintly green material, largely isotropic, but enclosing groups of tiny birefringent crystals, with individuals reaching about 0·01 mm. Examination in high power shows that the apparent individuals are themselves vermicular aggregates; since they have refractive index close to 1·565, it is suggested that they are referable to the kaolin group. Distinctly green material, picked from the hand-specimen, proved to be anisotropic only locally, and to have refractive index close to 1·60. Olivines in this rock are preserved in a yellowish-brown anisotropic material with high refringence, probably iddingsite.

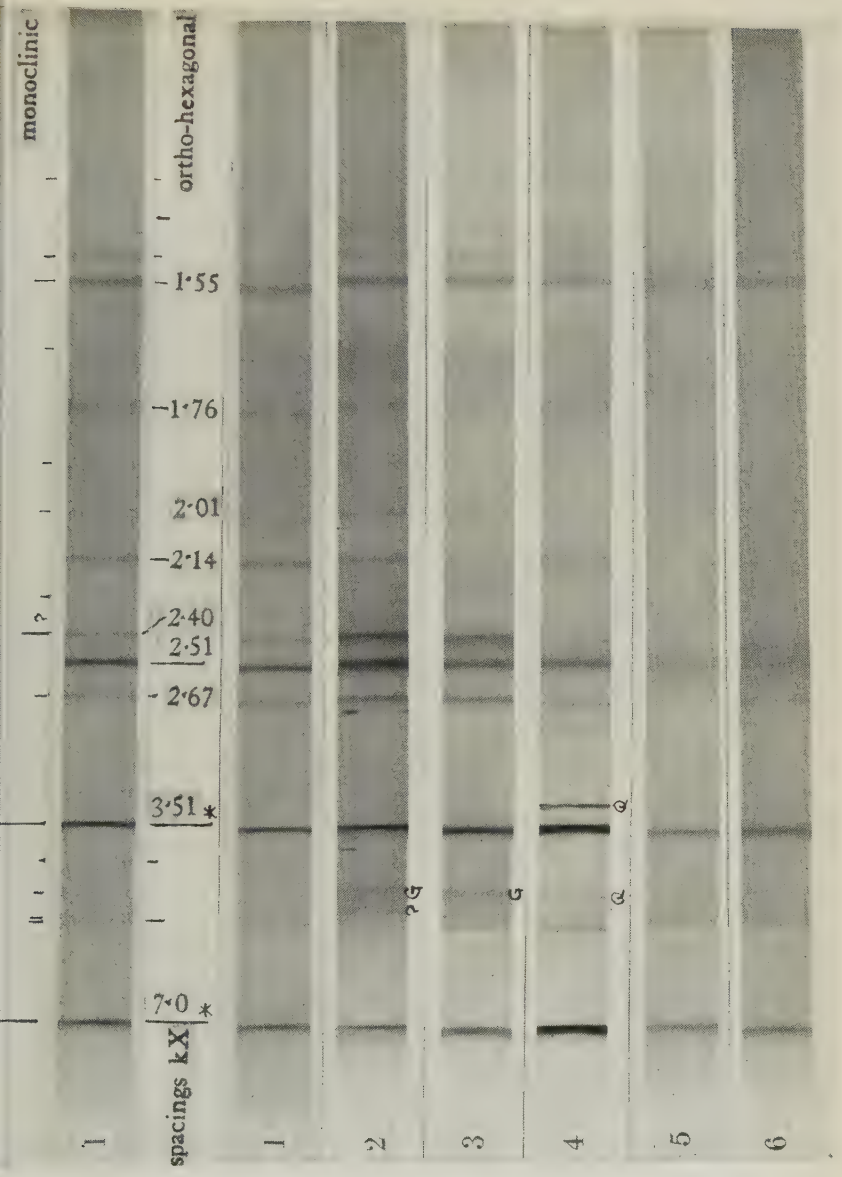
Material from Carnlough. (V. A. Eyles, B/0, Geol. Survey slice nos. N.I. 61 and N.I. 173.) Blackburn, 2·1 miles NNW. of Carnlough: 1-inch I. 14, 6-inch Antrim 25 NE.—Streaky brown, greenish-brown, and green rock, with no trace of basaltic texture remaining. At least three mineral phases are present: (i) fine-grained fibrous or vermicular mineral of the kaolin group, probably, since it appears to extinguish parallel, kaolinite; (ii) green apparently isotropic material having refractive index 1·645; (iii) opaque material, including iron oxides. In the brown parts of the rock, isotropic material with refractive index a little below 1·590 is present; this may be gibbsite.

The prevalence of apparently isotropic materials in these rocks suggests an extremely fine state of division if crystalline minerals are present (as the X-ray patterns indicate). Optical identification is not satisfactory under these conditions. The indications are, however, that a kaolin-group mineral, probably kaolinite, is developed in places sufficiently coarse to be anisotropic under the microscope. This is in addition to the green, more or less isotropic material with refractive index between 1·60 and 1·650 which is taken to be chamosite. Note also that all these rocks are contaminated by iron oxides and other minerals.

Petrographical data for specimens E. 20568 and B.M. 34121. By J. H. Taylor.

Specimen from Thingdon mine, Finedon, Northamptonshire. (E. 20568. Geol. Survey slice number).—The rock is a chamositic oolite from the upper chamosite–kaolinite group of the Northampton Sand (Taylor,[1]

[1] J. H. Taylor, Petrology of the Northampton Sand ironstone formation. Mem. Geol. Survey, 1949.



G. W. Brindley: X-Ray Photographs of Chamosite Minerals

949, pp. 15–16, 29–30; cf. pl. VI, fig. 1). This is the uppermost subdivision of the Northampton Sand, only locally present and resting nconformably on the underlying workable ironstone. The specimen ontains ooliths of green to yellow-green chamosite, 0·2–0·5 mm. across he longer diameter, set in a groundmass of fine-grained green chamoitic mud. Some of the ooliths are slightly distorted, some broken, while thers consist of a central pellet or fragment of chamosite surrounded ' green or yellow-green coatings. Many of the ooliths have thin zones r rims of a yellow-brown colour indicating partial oxidation. The hamosite mud of the groundmass is relatively fresh.

The sample selected for X-ray work consisted of a mixture of ooliths nd groundmass. Oxidized material was, as far as possible, avoided but the hamosite of the sample cannot be regarded as completely unaltered. The refractive index ranged 1·66–1·67.

Specimen from Kongens mine, Kongsberg, Norway. (British Museum, Natural History Collection, B.M. 34121; Geol. Survey slice no. F. 5405). —The specimen consists of green 'chloritic' material from a vein at Kongens mine and is of hydrothermal origin. In thin section it consists n part of relatively clear yellow-green to olive-green material of moderate birefringence occurring as minute flakes or very fine-grained vermicules vermicules are about 0·01 mm. across and up to 0·05 mm. long). Mixed vith this clear material are olive-brown to brown, opaque, finely granular patches resembling in appearance altered chamosite from the sedimentary iron-ores. The clearer material is optically negative and has refractive index of approximately 1·66, the value ranging upwards vith increasing density of colour.

Explanation of Plate XII.

X-ray powder diagrams of chamosites from following sources: (1) Ayrshire, cotland; (2) Knowehead, Co. Antrim; (3) Carnlough, Co. Antrim; (4) Wabana, Newfoundland; (5) Kongsberg, Norway; (6) Finedon, Northamptonshire.

Co-$K\alpha$ radiation. Camera radius, 100 mm. Powder plate set to record spacings 10 kX. Calculated diagrams for monoclinic and ortho-hexagonal structures are ompared with observed diagram for specimen (1); lines marked * are basal reflections which are enhanced in the photographs due to preferential orientation. Q ndicates lines due to quartz, G lines due to goethite, and ? lines not identified.

A serpentine mineral from Kennack Cove, Lizard, Cornwall.[1]

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Mineralogy Section, Building Research Station, Department of Scientific and Industrial Research.

[Read June 8, 1950.]

Introduction.—A white soapy mineral which could not easily be identified in the hand-specimen was collected from a vein in the serpentine rock at Kennack Cove, Lizard, in April 1949. From a preliminary optical examination the mineral was thought to be saponite, and therefore of interest to clay mineralogists. Chemical analysis, X-ray, thermal, and optical investigations were carried out at the Building Research Station.

Occurrence and optical data.—The mineral occurs in a vertical vein cutting the serpentine rock on the north-east side of Kennack Sands. The vein is about 2–3 inches wide and shows up very white against the dark serpentine rock. The mineral is pure white in colour, soapy to the touch, and soft (H. 2). It is an irregular aggregate of small crystals, which are for the most part hexagonal plates. In thin section the mineral is seen to have a distinct basal cleavage and a birefringence of about 0·01. Microscopical examination of powder specimens shows the plate-like habit of the grains, and also the presence of a second mineral which is probably talc. The plates, when lying flat, give a centred uniaxial negative interference figure. The mineral has ϵ 1·545 and ω 1·555, $\omega - \epsilon$ 0·01. These optical properties may be easily confused with those of saponite. Chemical analysis (table II) or X-ray examination (table III) will, however, distinguish between these two minerals. Table I gives the optical properties, with those of saponite, chrysotile, and antigorite for comparison.

Table I. Comparison of optical data.

	Serpentine mineral.	Saponite.	Chrysotile.	Antigorite.
ϵ	1·545	—	α 1·493–1·546	α 1·555–1·564
ω	1·555	1·555	β 1·504–1·550	β 1·562–1·573
$\omega - \epsilon$ ...	0·01	0·01 ±	γ 1·517–1·557	γ 1·562–1·573
2V ...	0°	0°	0–50°	20–90°
Sign ...	negative	negative	positive	negative

[1] Crown copyright reserved, published by permission of the Director.

Chemical analysis.—The analysis (table II) calculated on the basis of 10 oxygens gives the formula $Mg_{5\cdot6}(Si,Al)_4O_{10}(OH)_{7\cdot9}$. Making allowance for the talc impurity, it is probable that the mineral is analogous to either chrysotile or antigorite $Mg_6Si_4O_{10}(OH)_8$, and so should be regarded as a serpentine mineral.

TABLE II. Chemical analysis of serpentine mineral from Kennack, Lizard. (Analyst, L. J. Larner.)

SiO_2	...	...	44·49	CaO	...	...	0·03
Al_2O_3	...	...	2·26	TiO_2	...	...	0·03
MgO	...	...	40·27	Fe_2O_3	...	...	0·48
H_2O+	...	...	12·80				
							100·36

Thermal analysis.—A differential thermal analysis of a small sample was carred out by D. B. Honeyborne of the Building Research Station, and the resultant curve is reproduced as fig. 1. The curve is very like others obtained from some magnesium silicate minerals, for example antigorite.[1]

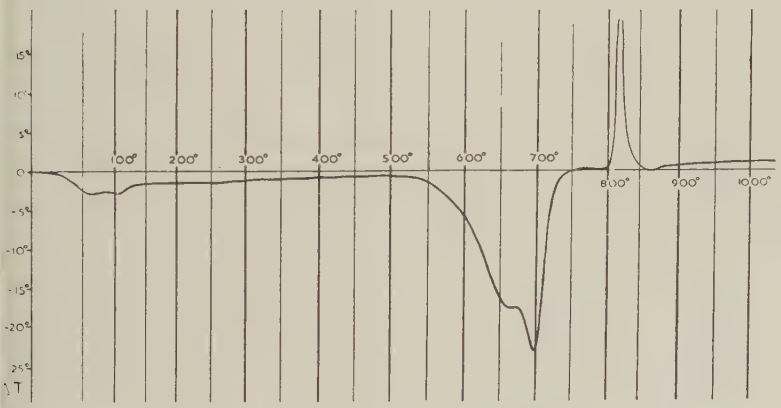

FIG. 1. Differential thermal analysis of serpentine mineral.

X-ray investigations.—Powder photographs and single-crystal rotation photographs of the serpentine mineral were taken. The powder photographs were taken on a 10-cm. camera using filtered Cu-$K\alpha$ radiation, λ 1·5374 Å. The measurements of spacings are given in table III. A powder photograph of a sample of saponite[2] from the Lizard (kindly

[1] S. Caillère and S. Hénin, Ann. Agronom. Paris, 1947, n. ser., vol. 17, p. 23. [M.A. 11–175.]

[2] This material is white or stained red, soft (H. 2) and soapy to the touch. It forms a granular mass of micro-crystals with refractive index about 1·555.

supplied by Dr. G. F. Walker) was also taken, and the spacings for this mineral are given for comparison, together with the spacings for chrysotile and antigorite as recorded by Selfridge.[1] It can be seen that there is no resemblance between the serpentine mineral and the saponite; and that there is some resemblance between the chrysotile, antigorite, and the serpentine mineral, but not complete agreement.

TABLE III. Comparison of X-ray powder spacings and intensities.

Serpentine mineral		Chrysotile*		Antigorite*		Saponite	
d_{hkl} Å.	*I.*	d_{hkl} Å.	*I.*	d_{hkl} Å.	*I.*	d_{hkl} Å.	*I.*
(9·4	5)†	—		—		18·8	10
7·51	10	7·364	9	7·355	8½	9·1	5
5·20	½	—		—		6·05	1
4·59	5	4·431	6	4·668	4	4·54	5
3·94	½	—		—		3·605	5
3·73	10	3·658	9½	3·641	9	3·00	4
3·105	3	—		—		—	
2·75	1	—		—		—	
2·635	1	—		—		2·605	6
2·580	1	2·571	8	—		—	
2·490	9	2·424	8	2·558	10	2·48	3
2·420	1	—		—		—	
2·29	½	—		—		2·26	2
2·205	½	—		—		—	
2·14	5	2·089	6	2·186	7	2·00	1
1·814	1	—		1·845	3	—	
1·780	3	—		1·794	4	—	
1·725	1	1·729	6	—		1·733	4
1·686	1	—		—		—	
1·657	½	—		—		—	
1·550	½	—		1·583	7½	—	
1·525	5	1·522	10	1·553	7½	1·533	7
1·497	5	—		—		—	
1·450	1	—		—		—	
1·407	½	—		—		—	
1·322	½	—		—		1·318	4
1·300	3	1·301	7	1·326	6	—	
1·268	3	—		1·273	4	1·268	2
1·238	½	—		—		—	
1·208	½	—		—		—	
1·159	2	1·187	½	1·160	2	—	
1·099	1	—		1·061	3½	—	
1·068	2	—		—		—	
1·051	2	1·039	2	—		—	

* Data from G. C. Selfridge, 1936 (loc. cit.). † Line due to talc impurity.

Since the mineral occurs as small flakes with distinct basal cleavage,

[1] G. C. Selfridge, Amer. Min., 1936, vol. 21, p. 463. [M.A. 6–476.]

it was possible to pick out individual flakes, one of which was mounted and used for rotation photographs. A 6-cm. single-crystal camera was used with filtered Cu-$K\alpha$ radiation, λ 1·5374 A. The flake was set up to give a, b, and c rotation photographs. The films produced are marred by the fact that some of the spots are extended into Debye powder lines. On indexing, however, it was found that the $hk0$ reflections were always spots, while hkl reflections were Debye powder lines. This suggests that there is preferred orientation of the layers.

The mineral was found to have a monoclinic unit cell with a 5·29, b 9·18, c 7·45 kX; β 91·4°; this cell leads to a calculated density of 2·52 g. c.c. The density determined by suspension in a bromoform-benzene mixture was 2·56 g./c.c.

In $h0l$ planes h is always even, in hkl planes h plus k is always even, indicating that the monoclinic cell is centred on 001, and that the probable space-group is $C2/m$, Cm, or $C2$.

The cell dimensions of chrysotile and antigorite have been given by Aruja[1] as:

Antigorite . . . a 43·39, b 9·238, c 7·265 kX; β 91·4°.
Chrysotile . . . a 5·32, b 9·2, c 14·62 kX; β 93·2°.

The serpentine mineral from Kennack has a smaller unit cell than antigorite, but since the ratio of the a-axes is about one-eighth and the other dimensions are similar it is possible that the two minerals have a similar structure. Using the kaolinite-like sheet structure suggested by Aruja for antigorite as a basis, the intensities of the various reflections were calculated. It was necessary to adjust the z-coordinates of the ions to allow for the difference between the c-axes of antigorite (7·265 kX) and serpentine mineral (7·45 kX). The calculated intensities for the reflections falling on the equators of the a and b rotation photographs are given in table IV, and it can be seen that there is reasonable agreement.

Conclusions.—The white mineral from Kennack with a formula of $Mg_6Si_4O_{10}(OH)_8$ has a monoclinic unit cell with a 5·29, b 9·18, c 7·45 kX, β 91·4°, probable space-group Cm, $C2$, or $C2/m$. It probably has a kaolinite-like sheet structure with all the octahedral positions filled with magnesium ions and so is the magnesium analogue of kaolinite. It is probably a variety of antigorite with very little iron substitution and thus has optical properties (uniaxial, negative ϵ 1·545, ω 1·555, $\omega - \epsilon$ 0·01) different from those usually associated with antigorite.

[1] E. Aruja, Ph.D. thesis, Cambridge, 1943; Min. Mag., 1945, vol. 27, p. 65.

TABLE IV.

Equator of a rotation photograph.

$4\sin^2\theta$ (calc.)	hkl.	F^2 (calc.)	Int. (obs.)	$4\sin^2\theta$ (obs.)
0·205	001	165·0	10	0·205
0·340	020	17·0	4	0·34
0·394	021	5·4	9	0·405
0·410	002	48		
0·533	022	7	—	—
0·615	003	18	3	0·615
0·680	040	1	1	0·680
0·704	023	3		
0·710	041	3	—	—
0·795	042	0	—	—
0·820	004	2	½	0·85
0·914	024	0	—	—
0·918	043	2	—	—
1·020	060	33	9	1·01
1·025	005	9		
1·040	061	6		
1·070	025	0	—	—
1·090	044	1	½	1·06
1·100	062	1	1	1·10
1·190	063	3		
1·230	006	3	½	1·27
1·233	045	0	—	—
1·27	026	0	—	—
1·325	064	0	—	—
1·360	080	1	—	—
1·400	046	0	—	—
1·435	007	1	—	—
1·470	027	0	—	—
1·58	047	0	—	—
1·64	008	0	—	—
1·67	028	0	—	—

Equator of b rotation photograph.

$4\sin^2\theta$ (calc.)	hkl.	F^2 (calc.)	Int. (obs.)	$4\sin^2\theta$ (obs.)
0·205	001	165	10	0·205
0·410	002	48	8	0·42
0·580	200	28	3	0·58
0·615	003	18	10	0·615
	201	75		
	$20\bar{1}$	39		
0·705	202	8	3	0·71
0·715	$20\bar{2}$	2		
0·82	004	2	1	0·84
0·83	203	1		
0·86	203	3	2	0·86
1·019	204	6	3	1·025
1·025	005	9		
1·040	$20\bar{4}$	7		
1·17	205	1	2	1·16
1·18	400	3		
1·19	$20\bar{5}$	3	3	1·19
	$40\bar{1}$	5		
1·20	$40\bar{1}$	9	2	1·20
1·23	006	3	½	1·23
	$40\bar{2}$	0	—	—
1·25	$40\bar{2}$	3	½	1·255
1·32	403	3	½	1·32
1·34	$40\bar{3}$	1	—	—
1·35	$20\bar{6}$	0	—	—
1·37	$20\bar{6}$	1	—	—
1·43	007	2	½	1·43
1·53	207	1	—	—
1·56	207	0	—	—
1·64	008	1	—	—
1·73	208	0	—	—
1·74	600	5	2	1·75
1·75	$20\bar{8}$	0	1	1·76
	601	2		
	$60\bar{1}$	1		
1·78	602	1	½	1·79
1·80	$60\bar{2}$	0	—	—
1·84	603	2	1	1·83
1·845	009	2	—	—
1·86	$60\bar{3}$	3	½	1·86

The Breece, New Mexico, meteoritic iron.

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[Read November 11, 1950.]

Introduction.

THE literature of meteoritics contains little information about the unusual Breece meteorite. This iron appears to be mentioned only in museum accession reports and catalogues of meteorite collections,[1] and the sole information supplied by these references consists of the approximate date of find (1921), the approximate weight (50 to 51 kg.), the approximate place of discovery, near Breece, McKinley County, New Mexico, in latitude 35° 18′ N., longitude 108° 18′ W., and the classification (Om). At the time of the discovery the hamlet of Breece[2] was the centre of a small lumbering industry, since moved SW. and SE. into more heavily forested areas, and today it has few inhabitants. The find consisted of a single well-preserved mass which was purchased in its entirety from an unnamed source by the Field Museum of Chicago approximately ten years after the iron was discovered.

The Heaston Collection of New Mexico meteorites at the University of New Mexico contains an interesting end-piece (fig. 1) weighing 920 grams sawn from the single mass which comprises the Breece find. The maximum dimensions of this specimen in three directions at right angles are $14{\cdot}5 \times 7{\cdot}1 \times 2{\cdot}3$ cm. This end-piece, registration number IOM–11 in the Institute of Meteoritics, the University of New Mexico, is the

[1] Ann. Rep. Field Mus. Nat. Hist. Chicago, 1932, for 1931, vol. 9, p. 133. H. H. Nininger, Our stone-pelted planet, Boston and New York, 1933, p. 158; The Nininger collection of meteorites. Mines Mag. Colorado School of Mines, 1933, vol. 23, pp. 6–9, 12 (p. 2 of reprint, 747 grams). [M.A. **5**-400, 405.] M. H. Hey, Second appendix to the catalogue of meteorites . . . British Museum. London, 1940, p. 24. [M.A. **7**-534.] A. L. Coulson, A catalogue of meteorites . . . Indian Museum. Mem. Geol. Surv. India, 1940, vol. 75, p. 57. [M.A. **8**-54.] S. A. Northrop, Minerals of New Mexico. Bull. Univ. New Mexico, 1942, no. 379 (Geol. Ser., vol. 6, no. 1), p. 186. [M.A. **8** 357, **9**-131.] S. H. Perry, The metallography of meteoric iron. Bull. U.S. Nat. Mus., 1944, no. 184, p. 167, pl. 47. [M.A. **9**–290.] F. C. Leonard, A catalog of provisional coördinate numbers for the meteoritic falls of the world. Univ. New Mexico Publ. Meteoritics, 1946, no. 1, p. 41. [M.A. **10**–171.] S. H. Perry, Meteorite collection of Stuart H. Perry . . . 1947, p. 3 (1487 grams, donated to University of Michigan).

[2] Breece is shown in the U.S. Geol. Surv. Topographic map of the state of New Mexico, edition of 1925, and on the U.S. Geological map of New Mexico, by N. H. Darton, edition of 1928; both maps 1 : 500,000.

subject of the mineralogical and chemical investigations reported in the present paper. This specimen, with its hackly fracture, presents evidence that the original 51-kg. mass was only a fragment of a larger mass which disrupted during transit through the atmosphere. Because of this

FIG. 1. Breece, New Mexico, siderite. A. Macro-etched surface: arrows point to cohenite lamellae. B. Exterior surface. $\times\frac{2}{3}$ (scale of inches).

evidence, an effort has been made to discover additional masses of the fall. Information supplied by Mr. Herbert Dick and Mr. L. E. Roderick of Albuquerque, New Mexico, and in particular by Mr. Charles Allen, a mineral collector of Trinidad, Colorado, leads to the apparently well-substantiated existence of a second fragment of the meteorite found in 1941 at one of the logging camps near Breece. Several searches with drag magnets and meteorite detectors have been made in the areas designated by these informants, but no additional finds were made. At the moment it must be presumed that the 51-kg. mass sold to the Field Museum constitutes the only fragment recovered and recognized from this fall.

Macroscopic examination.

The external crustal surface of the Breece meteorite (fig. 1B) is typical of most iron meteorites. The colour is dark brown to black, and the surface exhibits ample evidence of piezoglyphs, pressure markings produced during flight through the earth's atmosphere. The crustal coating ranges from vanishingly thin to a maximum thickness of 0·5 mm. The oxidation of the meteorite is limited to the surface and along cracks which penetrate into the interior of the iron (fig. 1A). Otherwise the meteorite is exceedingly well preserved; and, from this observation, it may be assumed that the fall did not antedate the find by many years.

The sawn surface of the meteorite was polished and macro-etched (fig. 1A). The metallographic structure is revealed to be that of an octahedrite. Lamellae of kamacite range in thickness from 0·6 to 1·2 mm., although some short lamellae have local thicknesses up to 1·8 mm. Because the greatest proportion of kamacite bands falls in the range of 0·6–1·0 mm., the Breece meteorite is classified in the division of siderites (Si), class octahedrites (O), subclass medium octahedrite (Om). The kamacite bands are in three systems forming good equilateral triangles; apparently the cut was made parallel to an octahedron plane. The kamacite bands are bordered by taenite lamellae which are, for the most part, much thinner than the kamacite. The angular interstices are filled with plessite fields, which, macroscopically, are triangular in form.

Another type of lamellae appears prominently in the macro-etch (fig. 1A, arrows). These lamellae are finely developed and either parallel to or perpendicular to kamacite bands. They average about 0·8 mm. in thickness and attain a maximum length of 1·8 cm. These lamellae in the Breece meteorite were noted by Perry (loc. cit., 1944) and referred to by him as Reichenbach lamellae. He says (p. 17): 'In octahedral irons

troilite is occasionally observed in the form of scattered plates or lamellae with a definite cubic orientation unrelated to the octahedral pattern.' Farrington says the Reichenbach lamellae form 'parallel to the faces of a cube or dodecahedron' (p. 100),[1] but contradicts this later (pp. 143, 150) by saying they form parallel to the cube. A similar arrangement of schreibersite parallel to planes of the dodecahedron is referred to by Farrington (pp. 100, 150) as the Brezina lamellae of Cohen. Dr. L. J. Spencer (personal communication), has pointed out that in the rhombic dodecahedron orientation traces of (110), (101), (011) and their opposite faces are parallel to octahedral edges, and traces of $(1\bar{1}0)$, $(\bar{1}01)$, $(0\bar{1}1)$ and their opposite faces are perpendicular to octahedral edges on the face (111), while cube traces are all parallel to octahedral edges. Inasmuch as these lamellae in the Breece meteorite are both parallel and perpendicular to the octahedral edges, they are presumed to be parallel to the rhombic dodecahedron. The lamellae figured by Perry (pl. 47), on a section of Breece approximately parallel to (111), are also parallel and perpendicular to octahedral edges; but when referring to them as Reichenbach lamellae (p. 167) no mention is made of either troilite or cohenite.

Several of these lamellae were isolated and subjected to three separate qualitative chemical analyses. In all three cases abundant carbon was obtained, and sulphur was completely absent. In two of these analyses phosphorus was absent, while the third showed a trace. The complete chemical analysis of the Breece siderite (table I) also shows phosphorus to be absent and confirms the presence of carbon, though the small amount of carbon would indicate that the analysed fragment contained at best but a small lamella. These analyses suggested that the lamellae are composed of cohenite. A quantitative analysis of another cohenite lamella is shown in table II. The results remove any possible doubts as to the nature of these lamellae. These same lamellae were observed in the microscopic examination of small polished sections. Immersion for 90 seconds in boiling alkaline sodium picrate, prepared by dissolving 12·5 g. of sodium hydroxide in 50 c.c. of water and adding 2 g. of picric acid, darkened the lamellae and confirmed that they are cohenite.

Cohenite is conventionally regarded as a rare meteoritical mineral. It has been reported from the Magura, Bendegó, Wichita County, Beaconsfield, Canyon Diablo, Cosby's Creek, and several other iron meteorites (Perry, loc. cit., 1944, p. 22). In physical appearance

[1] O. C. Farrington, Meteorites, their structure, composition and terrestrial relations. Chicago, 1915.

cohenite is identical with schreibersite, and distinguished from it with difficulty. The above test with boiling alkaline sodium picrate is the most diagnostic test for cohenite. In truth, it is to be expected that a more universal application of this etch test to polished sections would reveal that cohenite is more common than it is now believed to be.

The absence of sulphur in the lamellae makes the designation of Reichenbach lamellae inapplicable to this structure, and, as the presence of schreibersite appears doubtful at best, the term Brezina lamellae is equally inapplicable. These lamellae are similar physically to both the Reichenbach and Brezina lamellae, but the comparisons must be confined to structural relations only.

Microscopic examination.

Two slices of the Breece, New Mexico, siderite were mounted in bakelite, polished, etched with nital for 15 seconds, and studied with the metallographic microscope.

The microscope study shows that the external zone of alteration is extremely small, and that the interior of the meteorite, except along fractures, is free from alteration. From the unaltered appearance of the kamacite, taenite, and plessite, it may be assumed that the interior of the meteorite did not become appreciably heated during the flight through the atmosphere. Heating and the resultant fusion crust were confined to a thin outer zone.

The meteoritical components present are kamacite, taenite, plessite, cohenite, and troilite. The principal constituent is kamacite, the alpha nickel-iron alloy. In the macro-etched specimen the kamacite displays a characteristic sheen which changes as the surface is rotated. On the well-polished sections the kamacite is tin-white and occurs either as narrow, tapering lamellae bounded by very fine, fairly regular taenite and usually in a field of dense plessite (fig. 2); or it occurs as broad plates of two types: (1) apparently structureless, and (2) granular. Some of the latter type of kamacite show several sets of well-developed Neumann lines. Because Neumann lines can be obliterated in the laboratory upon heating, their presence is further confirmation that the siderite did not become heated in the interior.

Taenite, the gamma nickel-iron alloy, occurs usually as fine, fairly continuous bands bounding the kamacite lamellae (fig. 2) and forming the distinctive Widmanstetter pattern. Taenite also occurs as highly irregular patches. These patches range in appearance from entirely taenite, through taenite with some interior plessite (the 'fleckig' or

'spotted' taenite), to dense plessite with a border of taenite, the result of an incomplete separation of kamacite and taenite. In these taenite-plessite patches the plessite becomes increasingly denser towards the centre of the patch.

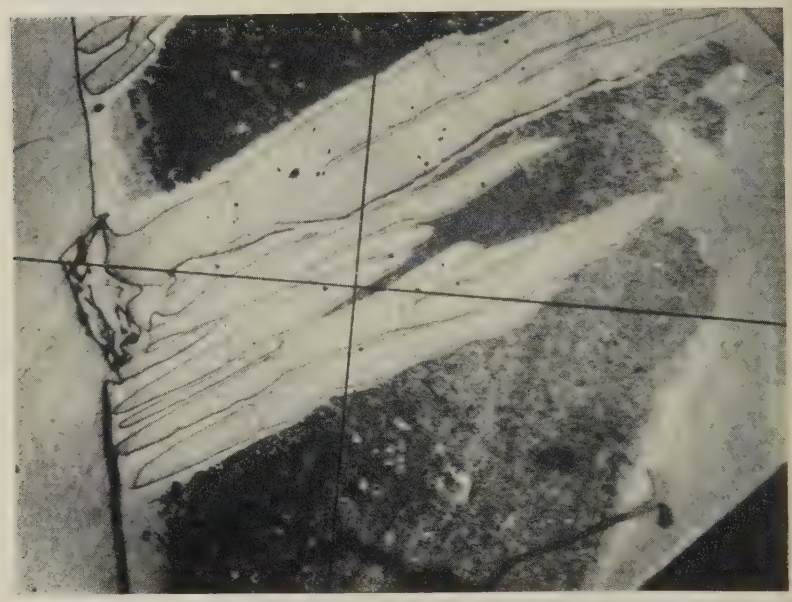

FIG. 2. Breece, New Mexico, siderite. Kamacite lamellae bounded by narrow taenite and surrounded by dense plessite. An irregular grain of brittle cohenite shows near the left-hand edge of the photomicrograph. ×125.

Most of the plessite on the polished sections is of the dense variety. Under high magnification the dense plessite can be seen to be composed of minute taenite particles embedded in a mesostasis of kamacite. These fields occupy the spaces formed by the lamellae of kamacite and taenite in octahedral arrangement.

In the polished sections cohenite generally appears as narrow, tin-white, brittle bands concentrated along the boundaries of contiguous kamacite plates, but occasionally an irregular grain of cohenite can be seen (fig. 2). That these are cohenite, and not schreibersite, was confirmed again by the boiling alkaline sodium picrate test.

Troilite occurs as tiny, irregular grains associated with the kamacite plates. In the Breece meteorite the troilite is typically brownish in colour.

Chemical analyses.

A portion of the Breece meteorite was removed from the mass, etched, and studied with a microscope to be certain it was representative of the meteorite. A chemical analysis was made by Dr. E. L. Martin, Department of Chemistry, the University of New Mexico, and the results are shown in table I. The analysis is in agreement with that of a normal medium octahedrite, though possibly a little higher in nickel than usual.

An analysis of one of the cohenite lamellae was made by Carl W. Beck. The results are shown in table II, and are consistent with the formula R_3C for cohenite where R = iron, nickel, and cobalt.

TABLE I. Chemical analysis made on 1·743 grams of the Breece siderite.

	%		
Fe	89·87	Specific gravity ...	7·93
Ni	9·26	Molecular ratio $\frac{Fe}{Ni}$ ...	10·20
Co	0·89		
S	0·11	Molecular ratio $\frac{Fe}{Ni+Co}$	9·34
C	0·03		
P	0·00		
Cl	0·00		
Cu	0·00		
	100·16		

TABLE II. Chemical analysis made on 0·4764 gram of cohenite.

	%	Molecular ratios.	
Fe	91·35	1·6356	
Ni	1·64	0·0279	3·054
Co	0·11	0·0019	
C	6·54	0·5450	1·000
	99·64		
Specific gravity	7·37		

Summary.—The investigation of the Breece, New Mexico, siderite shows it to be a medium octahedrite (Om) composed chiefly of kamacite with lesser amounts of taenite bands and plessite fields. The most interesting constituent is cohenite as well-defined lamellae in the unusual arrangement similar to the familiar Reichenbach and Brezina lamellae.

Jacobsite from the Tamworth district of New South Wales.

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[Taken as read November 2, 1950.]

TWO new occurrences of the rare manganese mineral jacobsite ($MnFe_2O_4$) have come to light in the course of mineragraphic studies carried out as part of the research programme of the Mineragraphic Section of the Commonwealth Scientific and Industrial Research Organization. The jacobsite occurs as a constituent of small bodies of high-grade manganese ore at Weabonga, near Danglemah, and at the Mount Sally mine, about 6 miles west of Danglemah, both in the Tamworth district of New South Wales.

The deposits occur in altered sediments, within a mile or two of a granite contact.[1] They are irregular lenticular veins ranging from a few inches to several feet in thickness, between altered slate walls. The veins do not exceed a length of 200–300 feet. The lode material consists of manganese oxides, chiefly psilomelane and pyrolusite, associated with quartz, rhodonite, and iron oxide. The manganese oxides are mainly supergene, and although the deposits are of high grade near the surface, it is doubtful whether they can be worked below the depths of 50–60 feet, owing to the increase in the amount of rhodonite and quartz relative to manganese oxides at this depth.

In the Weabonga ore the jacobsite occurs as narrow seams and lenticles, about 0·5 cm. across, and 3·0 cm. long enclosed in, and partly replaced by, pyrolusite and psilomelane. The seams consist of clusters of small crystals, which in the central parts of the seam show well-shaped octahedral form, with smooth facial contacts (fig. 1). Towards the margin of the seam, and where in contact with pyrolusite or psilomelane, the crystals are corroded and rounded (figs. 1 and 2), and at the margins of the seams small corroded grains of jacobsite are isolated as residuals in larger areas of pyrolusite and psilomelane (fig. 2).

The range in grain-size of the jacobsite crystals is from about 0·15 mm. across (large unreplaced grains) to about 0·01 mm. across (small residuals). Away from the immediate edge of the seams and clusters of

[1] C. L. Knight, Mineral resources of Australia, Summary Rept. no. 7, Manganese, Bureau Min. Res., Geol. Geophys., Australia, 1947, p. 13.

jacobsite the pyrolusite and psilomelane are free from inclusions of this mineral.

The Mount Sally occurrence is similar. Here the jacobsite seams are 1·0–2·0 cm. long and 0·2–0·3 cm. wide, with the individual octahedra of jacobsite ranging from 0·025 to 0·055 mm. across. Psilomelane is the dominant enclosing mineral, and the proportion of jacobsite seams is rather less than in the Weabonga ore.

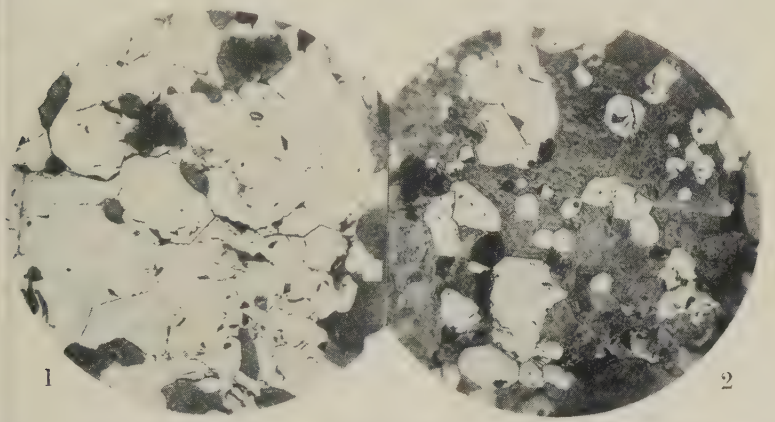

Fig. 1. Centre of a jacobsite seam, showing the crystal outlines of jacobsite grains (white outlined by black lines) where they are in contact with one another, and their embayed character where they are in contact with psilomelane and pyrolusite (black). Etched with $SnCl_2$. ×60.

Fig. 2. Marginal portion of the same seam of jacobsite as in fig. 1, showing more advanced replacement of the jacobsite (light) by the psilomelane and pyrolusite (black). Etched with $SnCl_2$. ×60.

The jacobsite is strongly magnetic. It takes a polish readily, and in polished sections exhibits its characteristic grey colour with an olive-yellow tint, which distinguishes it from magnetite or other strongly magnetic minerals. It is isotropic, and is negative to all standard etching reagents, but tends to be slightly tarnished brown with stannous chloride. In all these respects it exactly resembles a specimen (779*a*) of jacobsite presented by Dr. J. A. Dunn, from the Beldongri mine, in the Central Provinces of India,[1] except that the yellow tint is slightly less pronounced. It is harder than the pyrolusite and psilomelane, and this

[1] J. A. Dunn, A study of some microscopical aspects of Indian manganese-ores. Trans. Nat. Inst. Sci. India, 1936, vol. 1, no. 7, pp. 103–124 (p. 108). [M.A. 7–157.]

combined with its crystalline form permits it to be broken free from the other minerals, when it can be concentrated magnetically.

Several grams of the Weabonga jacobsite were separated with a hand magnet from crushed ore, and passed through a 100-mesh sieve. The minus 100-mesh fraction was cleaned as far as possible with a magnet, leached with dilute HCl, and dried at 60° C. A portion of the final product was mounted in a bakelite briquette and polished. Examination under the ore microscope revealed that it consisted essentially of jacobsite and psilomelane, with some finely microcrystalline pyrolusite intergrown with the psilomelane. A micrometric analysis of the opaque constituents gave jacobsite 91·4, psilomelane (and pyrolusite) 8·6 volume per cent. The differences in specific gravity of these minerals are so slight that these figures may be taken as weight percentages.

A chemical analysis of the powder has been made by Mr. G. C. Carlos in the C.S.I.R.O. Mineragraphic Laboratory, with the following result:

Fe_2O_3.	FeO.	MnO.	MgO.	CaO.	insol.	H_2O.	Total.
62·67	nil	32·97	0·08	nil	0·19	3·70	99·61

Allowance can be made, within narrow limits of error, for the 8·6 % of psilomelane (plus pyrolusite) in the analysed powder by assuming (1) that all the impurity was psilomelane and (2) that its manganese content is similar to the ideal composition of psilomelane given in Dana's System of mineralogy, 7th edit., 1944, vol. 1, p. 669. Psilomelane on this basis contains 7·4 % MnO, and 72·8 % MnO_2:

$$\frac{7{\cdot}4}{100} \times 8{\cdot}6 = 0{\cdot}64 \text{ MnO}$$

$$\frac{72{\cdot}8}{100} \times 8{\cdot}6 = 6{\cdot}26 \text{ } MnO_2\text{, equivalent to } 5{\cdot}11 \text{ MnO.}$$

The total MnO contributed by the 8·6 % of psilomelane to the chemical analysis is thus indicated as 5·75 %, leaving for jacobsite Fe_2O_3 62·67, MnO 27·22, MgO 0·08.

Assuming that the H_2O of the analysis belongs to the psilomelane or is otherwise adventitious, and recalculating to 100 %, gives the composition of the jacobsite as that shown in column 1 below.

It is thus apparent that the composition of the Weabonga jacobsite, like the jacobsite of Nordmark,[1] corresponds closely to that of the theoretical $MnFe_2O_4$; and that it contains notably less magnesium than the jacobsite from Jakobsberg, the type locality. In this respect it

[1] L. J. Igelström, Sur la jacobsite de Nordmark et sur les jacobsites en général. Bull. Soc. Franç. Min., 1887, vol. 10, p. 185.

resembles the Indian jacobsite, but is otherwise different in that it contains more iron and less manganese. Since much of the jacobsite in the Beldongri mine is intimately intergrown with ex-solution lamellae of hausmannite, it seems likely that the analysed specimen, which was optically homogeneous, still retained some hausmannite in solid solution, whereas the Weabonga jacobsite does not.

	1.	2.	3.	4.	5.
Fe_2O_3	69·7	69·24	69·4	68·25	59·31
FeO	nil	—	—	—	—
Mn_2O_3	—	—	—	4·03	8·36
MnO	30·2	30·76	30·6	20·72	32·20
MgO	0·1	—	—	6·41	0·13

1. Jacobsite from Weabonga, New South Wales.
2. Composition of $MnFe_2O_4$.
3. Jacobsite from Nordmark, Sweden (Igelström, 1887).
4. Jacobsite from Jakobsberg, Sweden (Dana's System of mineralogy, 6th edit., 1892, p. 227.
5. Jacobsite (779*a*) from Beldongri mine, Central Provinces, India. Calculated from the composition given by Dunn (loc. cit., p. 110).

A jacobsite so closely corresponding to the ideal composition $MnFe_2O_4$ is of interest in view of the statement by Mason[1] that although jacobsite has generally been considered as manganese ferrite, none of the published analyses of jacobsite corresponds to the theoretical $MnFe_2O_4$.

[1] B. Mason, Mineralogical aspects of the system FeO–Fe_2O_3–MnO–Mn_2O_3. Geol. För. Förh. Stockholm, 1943, vol. 65, pp. 97–180 (p. 157). [M.A. 9–32.]

On a cupriferous Lewisian para-gneiss.

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THE writer has recently been engaged in the investigation of a 'Lewisian' inlier within the Moine series of Glen Dessarry, at the head of Loch Arkaig, Inverness-shire. The inlier is largely composed of a series of hornblendic and felspathic gneisses but includes small lenses of marble and calc-silicate rocks. All these are to be reported later in two separate communications. There are in addition, however, some steeply inclined and impersistent ribs of mica-schist, granulite, and diopside-epidote-rocks, which will only be incidentally mentioned in those accounts in so far as they bear upon certain specific problems. In themselves they are apparently of little petrographic interest except for the occurrence of an unusual copper-bearing meta-sediment. This it seems convenient to record in a separate note.

The exposures of this rock occur in the lower slopes of the hill on the south side of Glen Dessarry, and near the head of the glen. The precise locality lies about 250 yards SSW. of the shepherd's cottage Cuil (pronounced 'cool'; on some maps A' Chùil) and about 100 yards west of the Allt Coire a' Bhoganaich. Here, within ultra-basic and basic hornblendic-felspathic gneisses there occur several thin, almost uncontorted impersistent sheets of foliated meta-sediments. These ribs, a few feet thick and sporadically exposed, extend for about 100 yards uphill to the south along the strike. The dip is westerly, almost vertical. The meta-sediments are chiefly semi-pelitic schists and granulites with thin pelitic-seams and with interbanded green diopside-rock also containing biotitic layers. The diopside-rock may contain scattered crystals of diopsidic augite by accession of which it gradually passes into bodies composed of such crystals. These are usually partly replaced by deep-green hornblende to form rock types from which felspathic gneisses are seen to derive by a process of pegmatitization.

In this rock association certain semi-pelitic ribs (unfortunately poorly exposed) carry small grains of a green mineral with the general appearance of malachite. These measure up to the size of a pin's head and are dis-

eminated throughout the rock. Perhaps about a dozen grains may be een with the naked eye in one square inch of rock. Under a pocket-lens hey show ragged margins interlocking with the parent rock. One or wo grains contain cores of a dark opaque mineral resembling bornite. A secondary origin is therefore indicated for the green mineral.

A thin section of the rock shows a medium- to fine-grained granoblastic ground of potash-felspar and oligoclase. Biotite is abundant as traight plates and almost black in the position of maximum absorption. There are no signs of crushing. Sphene, quartz, magnetite, and apatite re accessory, together with a translucent grass-green mineral in regular-shaped patches about 0·1–0·5 mm. across, or occasionally forming straight-sided laths within, and along the cleavage of, biotite. These orms are all cryptocrystalline, sometimes with fibrous habit. The green nineral shows aggregate polarization. Absorption masks the interference tints. It is not pleochroic. In reflected light it is of apple-green olour with a dull, earthy lustre.

The green mineral was carefully picked out of the rock and a clean ample separated under the binocular microscope. By kind permission f Professor W. Q. Kennedy an X-ray powder photograph of the sample as taken in the Geology Department, Leeds University. The film natched a standard powder photograph for malachite.

A copper estimation performed on the parent gneiss gave the following result: copper 0·24 % (analyst Herdsman).

It may here be of interest to compare this figure with estimations by teiger, quoted by Siebenthal,[1] carried out on a composite of 235 amples of Mississippi delta mud. This composite material is shown by ettijohn[2] to be about $\frac{1}{3}$ sand, $\frac{2}{3}$ silt and clay. Steiger's analysis, mongst other values totalling 100·62, gave SiO_2 69·96, Al_2O_3 10·52, uO 0·0043 %.

The problem of the origin of the Glen Dessarry copper arises. There ppears to be no vein mineralization and the copper-bearing gneiss is ot smashed to provide easy passage for hydrothermal solutions. It has lso resisted mechanical penetration by pegmatite, a penetration greatly ffecting the associated ultra-basics. However, both from its intimate ssociation with the last, and from its petrography, it seems to have been netasomatically modified by pegmatite. Perhaps then the copper was oncomitantly introduced. One fact in opposition to this, however, that pegmatitization dominates the whole inlier, whilst the copper-

[1] C. E. Siebenthal, Bull. U.S. Geol. Surv., 1915, no. 606, p. 72.

[2] F. J. Pettijohn, Sedimentary rocks. New York, 1949, p. 271.

bearing rock seems to be extremely localized, no other cupriferous material having yet been detected within the inlier.

In view of the complex history of the district, the poor exposures of the cupriferous rock, and the unknown role of meteoric waters, no definite conclusions are advanced regarding the origin of the copper. Nevertheless it is tentatively suggested that a syngenetic, sedimentary origin may be worthy of consideration.

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